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CYCLOGENESIS IN THE LEE OF THE CANADIAN ROCKY MOUNTAINS

BY

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled "Cyclogenesis in the Lee of the Canadian Rocky Mountains", submitted by Yong-Seung Chung in partial fulfilment of the requirements for the degree of Master of Science.

NIMBUS 4 APT Satellite picture (1849Z 10 March 1972) showing cyclogenesis (marked ⊕) not associated with an upper trough and resulting clouds in the lee of the SW Alberta Range. The cyclone is initiating under a 500-mb strong cross-barrier flow over the Range. There also appears to be strong subsidence (cloud free region) from the Divide to about 50-100 km to the east. The banded high cloud shield over the Mackenzie Mountains is associated with a jet stream. White dotted lines are 500-mb contours.

ABSTRACT

The lee of the Alberta Rockies has long been known as a major cyclogenetic region. The purpose of this study is to examine lee cyclogenesis in the Canadian Rocky Mountains and to test whether the present theories of cyclone development can be applied.

After a review of important classical and recent work on cyclogenesis and orographic barrier flow, results are presented of a study of lee cyclones and related synoptic features associated with the Canadian Cordillera. Examination of the synoptic maps of Western Canada for 1958 revealed 146 cases of lee cyclone activity. These cyclones were classified according to intensity, based on Schallert's 1962 system of cyclonic types. About half the cyclones were found to be of Type A, the most intense and numerous. The remainder comprised three types of small, local cyclones, the usually weak and non-developing lows of Type B, C and D.

Detailed analysis indicates that the frequency maximum of lee cyclogenesis in the Canadian Rockies is not single, but composed of three distinct centres. The main centre of activity is located in the lee of the Southwestern Alberta Range, while two secondary maxima are found in the lee of the Northern B.C. Range and the Mackenzie Mountains. The frequency maxima appear about 200-250 km from the Continental Divide. This distance may be identified with a region of vertical stretching and, perhaps, be a parameter important in synoptic-scale mountain wave theories, but the evidence so far is inconclusive.

In the development of pressure systems over mountains, a divergence field is predominant over the windward side which, accompanied by orographic ascent, produces a cell of divergence in the lower and mid-troposphere, and a compensating cell of convergence in the stratosphere and upper troposphere. In the lee of the range, the cyclogenetic fields are generated and compensated in the approximately converse manner necessary for the development of extratropical cyclones, as postulated by Dines, Scherhag, J. Bjerknes and Sutcliffe.

Using a smoothed topography, orographically induced vertical motions were computed. Vertical velocities of about 3 to 7 μb/sec were found to be associated with typical cases of cyclogenesis. Vorticity advection in the middle and upper troposphere can result in cyclogenesis, but the basic mechanism responsible for the initiation of most lee cyclones, particularly of feeble and non-developing systems, seems to be mid-level divergence. This condition can be usually identified with characteristic flow patterns aloft. In particular, it was observed that most lee cyclones formed initially under the eastern margin of a 500-mb, orographically intensified, diffluent cross-barrier flow, superimposed on a zone of low-level convergence and orographic descent. However, sudden deepening of lee cyclones was found to be associated with upper cold troughs and positive vorticity advection, as required by Petterssen's hypothesis.

DEDICATION

To My Loving Parents

Whose Encouragement, Support, Understanding

and Strong Belief in Education

Have Made This Work Possible.

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CHAPTER I

HISTORICAL RETROSPECT

Showers occur in nature more readily than we can account for by laboratory or theoretical calculations.

B. J. Mason[1]

1.1 Introduction

The development and behaviour of weather systems interacting with mountains is a challenging subject for synoptic analysis and research. Orography plays an important role in the atmospheric circulation, for mountain ranges generate and modify weather systems on all scales of atmospheric motion. Thus, for example, the lee of the Rocky Mountains is a favoured cyclogenic region, with the Alberta and Colorado lee regions known to be particularly important.

Though the behaviour of cyclones in mountainous regions is still not fully understood in several respects, some of the effects of orography on the synoptic and planetary scale motions of the atmosphere have been introduced into the numerical weather prediction models of recent years. Diverse scale motion systems, such as cyclogenesis, air flow over mountain ranges and lee waves, have been modelled also in the laboratory by Fultz (1952), Fultz and Kaylor (1959), and by Long (1959).

[1]Speaking at the Symposium held on the occasion of the First Centenary of the Canadian Atmospheric Environment Service in Toronto, October 26 to 28, 1971, as quoted in the Edmonton Journal.

The purpose of this study is to review briefly the more important theories of cyclone development and intensification, and to investigate some aspects of the behaviour of cyclones generated in the lee of the Canadian Rocky Mountains. It is hoped that the results may be useful in forecasting.

1.2 Cyclone Models and Polar-Front Theory

The first significant advances in theoretical and synoptic meteorology were made about 1820 and continued along several frontiers. Excellent historical reviews of these early developments are given by Petterssen (1956, 1969), Bergeron (1959), and Palmén and Newton (1969).

A remarkable but much forgotten cyclone model was published by Jinman in 1861. It shows the pattern of streamlines of a cyclone with quite realistic wind directions and speeds. A strikingly modern feature of the model is its two lines of discontinuity, the first known reference to the existence of fronts[1].

After a world voyage FitzRoy, in 1863, drew a realistic picture of four cyclones on the boundary between tropical and polar air masses[2]. Other schematic models were constructed later, one by Ley in 1878 which shows the three-dimensional structure of a frontal cyclone, and another by Abercromby in 1883, a low pressure system with oval isobars[3]. These two models give some of the essential features of cyclones, such as

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[1]See Petterssen (1969) page 11.

[2]See Petterssen (1956) page 214.

[3]See Bergeron (1959) page 448.

their direction of movement, the cloud and precipitation patterns, and the wind directions. Bigelow noticed in 1902 that centres of cyclones were found on lines separating warm and cold currents[1].

There is, furthermore, an interesting diagram by Van Bebber published in 1890. This shows a sea-level centre of low pressure with counter-clockwise, convergent airflow near the surface, while the flow aloft consists of a divergent flow[2]. This evidently gives the principal features of cyclone structure and is in excellent agreement with the depressions described by Dines (1919).

Petterssen notes that Shaw, in 1911, was able to modify slightly Abercromby's model with the aid of detailed synoptic observations. In a study on the processes of development of circular storms, Shaw (1906) also discussed broadly their trajectories and the formation of secondary storms behind a primary depression from the analysis of weather maps for 1882. He defined pressure troughs, and found northerly to westerly gales with a very cold descending current in the west and southwest region of a cyclonic area, while in the southern region he observed steadily decreasing pressures with a warm moist southerly flow. But weather-fronts in the presently accepted sense were not discovered, and the causes of the pressure changes accompanying cyclone development were not understood at that time.

The upper air motions associated with weather and the discontinuity of weather phenomena in cyclones were recognized early

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[1]See Brunt (1939) page 307.

[2]Van Bebber also includes a sketch of an anticyclone. See Petterssen (1969) page 13.

by some investigators such as Ley and Köppen[1] in 1882, but the most striking and realistic cyclone model was developed by J. Bjerknes (1918) from a careful analysis of numerous synoptic weather maps. This model clearly showed the formation of an extratropical cyclone on a surface of discontinuity. A little later this model was modified slightly by J. Bjerknes and Solberg (1921). A prominent aspect of this work was the discovery of the *steering line*, now called a warm front[2].

J. Bjerknes found from the analysis of streamlines that cyclones did originate on the shear zone named later a *front*, a line of separation between polar and tropical air currents. An important discovery in the search for sufficient causes for cyclogenesis is that there are links in the interchange of air between the cold polar region and the equatorial zone along the line of discontinuity.

Furthermore, Bjerknes provided a realistic explanation of the weather phenomena associated with the model cyclone, and described several other remarkable features such as the coupled systems of cyclone families whose members are in various stages of their life cycle, frequently attaining great intensities during the seasons of strong horizontal temperature gradient.

Hypotheses concerning the fundamental mechanism of cyclone development, and the empirical results of the *polar front* investigations were strongly supported by the theoretical studies of V. Bjerknes (1921). These discoveries eventually ushered in a new era of synoptic meteorology, and revolutionized forecasting for years to come.

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[1] See Bergeron (1959) page 449.

[2] The existence of a cold front was then already known to meteorologists such as Köppen as a *squall line*. See Bergeron (1959) page 449.

1.3 Theoretical Developments

Petterssen (1956) points out that Helmholtz, as early as 1858, derived the vorticity theorem which only much later was used to elucidate the development mechanism, and later still was applied successfully in synoptic forecasting. Helmholtz (1888) also introduced the concept of dynamic instability of the internal motion in isentropic surfaces, and showed that two air currents of different density, temperature and velocity could flow side by side in a stable condition, separated by a surface of discontinuity and that, moreover, the perturbations on such a shear zone might become unstable and grow under certain conditions into waves and vortices. It was later recognized that such a fundamental principle may be applied to the initial stage of cyclone development and to gravitational meso-scale waves, even though the Helmholtzian waves are only of the order of one kilometer in wave-length. An expression for the slope of a similar kind of discontinuity surface was first derived by Margules (1906) using geostrophic winds, gravitational and Coriolis forces, and air densities in the warm and cold air.

Margules (1903) was also the first to elucidate the fundamental energy problem of a cyclone in a theoretical study, showing how the kinetic energy of a cyclone may be derived from the available potential energy associated with horizontal temperature contrasts. In later years J. Bjerknes (1918), V. Bjerknes (1921), and J. Bjerknes and and Solberg (1921, 1922) added to these ideas with their theory of the polar-front by demonstrating that the essential feature of the life cycle of cyclones is the concurrent energy conversion. They recognized that the kinetic energy necessary for cyclogenesis is obtained from the

appreciable air-mass contrasts across the discontinuity surface between the polar and equatorial source air masses. Brunt (1930) also called attention to Exner's hypothesis that cyclones form at the edge of a discontinuous outbreak of cold air, an apparent modification of the idea of V. Bjerknes (referred to by Shaw (1920)) that cyclolysis results from the cutting off of the supply of warm air by the protrusion of two branches of the polar current.

Circulation theorems for the *baroclinic state* were developed by V. Bjerknes (1911, 1921) as early as 1898. V. Bjerknes and collaborators (1911) described cold and warm waves with a convergence line prior to the discovery of the polar-front proper. The wave theory of the origin of extratropical cyclones was primarily developed by V. Bjerknes (1921), though there existed the Helmholtzian perturbation concept. He also showed that *the general surface of discontinuity is a phenomenon of baroclinic character, and that inertia waves are due to gravity, the sloping internal surface of discontinuity, and the earth's rotation.* A quasi-elliptical low pressure system, an idealized theoretical wave, was applied to the practical problem of a cyclone as a type of wave on isobaric surfaces, with a wave-length of 2000 km and a period of 36 hours. The empirical facts gave increasing evidence for the theory of the circular vortex that cyclones could be considered as waves of diverse complexity and variety.

Thus evolved gradually a schematic picture of the general circulation with families of cyclones along the polar-front. The formation of new cyclones likely took place where the front had become stationary, but the cyclones propagated as the disturbance

increased in amplitude. Though the tilting of the discontinuity surface of the perturbation motion can generate unstable waves, which are the inertia waves formulated by V. Bjerknes (1921), it became clear that the effects of inertia, gravitational and shearing instability alone could not be considered adequate to explain fully the problem of cyclogenesis discussed by Haurwitz (1941).

Only little was known of the upper air structure at the end of the 19th century, but with the aid of early balloon soundings, it was demonstrated by V. Bjerknes and collaborators (1910) and by Dines (1912) that cyclones in general consist of relatively cold air aloft. The movements of cyclones were also known to be in the general direction of the upper air flow or of the mean flow in the warm sector, in accordance with the original ideas of the Norwegian and German-Austrian schools, but further studies had indicated that the detailed motion of such systems was often very complex. In later years it was realized that the individual cyclones move forward relative to the long waves and are effectively steered by the broad-scale flow pattern aloft.

Margules (1904) enunciated that *a small net divergence or convergence is a measure of the surface pressure change.* V. Bjerknes and collaborators (1911) also suggested the occurrence of a line of convergence at the ground in a cyclone when the upper layer had a sufficiently strong motion, and it likely appeared a little before a kinematically unstable advancing upper wave. One could then expect a fall of pressure and temperature with the approach of such a convergence area.

Following these early findings, Dines (1919) introduced the

idea of *a compensation in the divergence and convergence fields in the lower and upper troposphere*, a very important mechanism of cyclone development. All these ideas finally culminated in the remarkable development theories of Brunt (1930, 1939), Scherhag (1934, 1937), J. Bjerknes (1937), J. Bjerknes and Holmboe (1944), Sutcliffe (1939, 1947, 1950), and Petterssen (1950, 1955). However, they all concluded that the principal reason for the falling pressure is that the air is removed horizontally and vertically from a centre of low pressure, a fundamental idea expounded originally by Margules in 1904.

Further insight into the problem was provided by Scherhag (1934, 1937)[1] who formulated the divergence theory of low-level development, with the postulate that *divergent upper winds must produce in general a fall of pressure (at the earth's surface) if they are not compensated by strong convergence below.* More specifically, Scherhag emphasized that *divergence in the upper troposphere is both necessary and sufficient for cyclonic development.* This evidently suggests that *cyclones should develop most frequently where net mass divergence readily occurs, such as in regions favoured by divergent flow aloft.*

1.4 The Margules-Bjerknes Development Scheme

J. Bjerknes (1937, and with Holmboe, 1944) introduced a physical relationship showing that surface cyclones are in intimate association with the flow patterns of the upper atmosphere, such that there is, in general, upper level divergence downstream on the east side, and convergence upstream on the west side of troughs in the westerlies.

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[1]See also Sutcliffe (1939).

Bjerknes interpreted the principle embodied in the tendency equation, which was derived earlier by Margules (1904), that surface pressure changes are due to the integrated effect of the horizontal mass divergence. With symbols defined on page 144, the tendency equation is derived from the hydrostatic and continuity equations:

$$\left(\frac{\partial p}{\partial t}\right)_z = -g \int_z^\infty \mathrm{div}_h \, (\rho \vec{V}_h) \, dz + (g\rho w)_z \tag{1}$$

It may be seen that the local pressure change at a certain level z is determined by the horizontal mass divergence at all levels above z, and by the vertical motion of the air through the base of the column at height z.

Now, for a level at the earth's surface the vertical motion is zero, and the pressure tendency equation becomes

$$\left(\frac{\partial p}{\partial t}\right)_o = -g \int_o^\infty \mathrm{div}_h \, (\rho \vec{V}_h) \, dz \tag{2}$$

Hence, it is clear that *the upper level mass divergence is the sole measure of the low level pressure changes over flat terrain.*

Bjerknes (1951) concluded that *we may classify their formation (of extratropical cyclones) as being due either to unstable frontal wave action or to unstable growth of an upper wave trough.* In practice it is found that both processes go on simultaneously in most intense cyclones. As it will be seen in Fig. 1, Bjerknes also explained how the deepening could be attributed to the relative horizontal displacement of the upper air wave with respect to the surface cyclone, and pointed out the role of the baroclinicity and the vertical shear. Evidently, a phase lag between the sea-level cyclone and the upper-level trough is necessary for the deepening of cyclones. These ideas are

essentially implicit in the perturbation wave theory of Charney (1947) which requires that the maximum horizontal mass convergence in the low level is taking place between the upper trough and the preceding ridge. New insight came also with the planetary-scale wave theory based on the principle of the conservation of vorticity, as developed extensively by Rossby (1939, 1940), and by Haurwitz (1941).

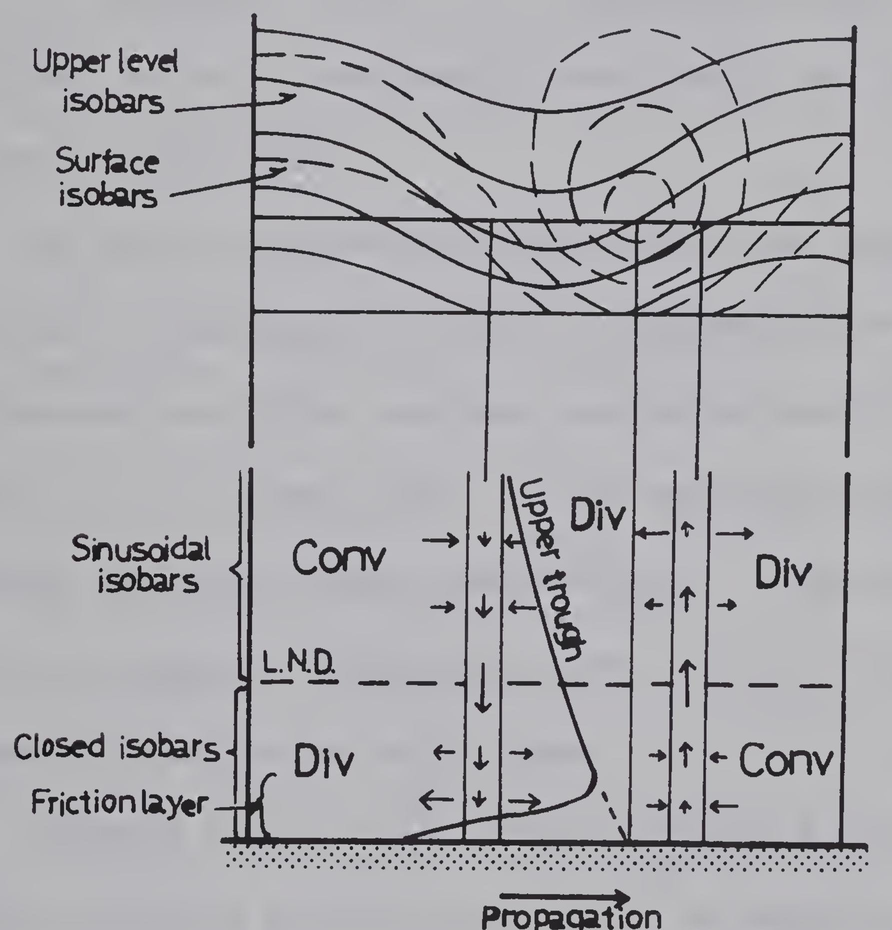

Fig. 1. A schematic distribution of divergence and convergence fields and the associated vertical motions in a typical cyclone. The phase lag between the upper trough and the sea-level cyclone is seen in the superimposed upper and surface maps. After J. Bjerknes and Holmboe (1944).

1.5 The Importance of Vertical Motions

In the previous section it was noted that the development of low-level disturbances is largely determined by the upper-level horizontal divergence fields. However, closely associated with the divergence is the equally important process of vertical motion which, though typically two orders of magnitude smaller than the horizontal wind, is crucial in the development of cyclones. As shown by Margules (1904) and Petterssen (1956), the numerical values of both the divergence and vertical velocity are usually very small and their accurate evaluation is difficult.

In early years such factors were also emphasized by V. Bjerknes and collaborators (1911) as *diagnostic and prognostic use of the solenoidal condition* and they evaluated vertical motions and divergence in a cyclone centre - both important measures of development of pressure systems in modern meteorology. Large values of vertical velocity and negative divergence were found in certain zones, which were later identified as polar-front.

Fleagle (1947, 1948) while estimating vertical motions in travelling pressure centres found that the magnitudes were highest in the middle troposphere near the 600-mb level. Moreover, strong mass divergence occurred in the upper levels of a cyclonic system, and the greatest low-level convergence took place in advance of a region of divergence aloft. Such studies provided the characteristic structure of a cyclone, and also conformed to the classical ideas such as of Dines (1919).

An important relationship between vertical motion and horizontal divergence was formulated by Sutcliffe and Godart (1942):

$$D_p = -\frac{\partial}{\partial p}\frac{dp}{dt} \tag{3}$$

The equation evidently states that the horizontal mass divergence is largely dependent upon the variation of the vertical velocity, and hence the vertical motion is also seen to be a fundamental mechanism in the development of pressure systems.

In a significant series of papers Sutcliffe (1939, 1947, and with Forsdyke; 1950) formulated a development scheme which stressed the essential importance of vertical shear in the development of cyclones. Sutcliffe applied *the vorticity changes in lower and upper levels* to the diagnosis of divergence, which correspond to the magnitude of the compensating divergences in the respective levels. Development may thus be explained in terms of the geometry of the thickness patterns, with the 1000-500 mb thermal wind field indicating the vertical variation of vorticity transport.

Attention was directed to the development fields associated with thermal troughs and ridges, and also with confluence and diffluence patterns. Sutcliffe and Forsdyke (1950) formulated a development equation where the sea-level divergence was considered in terms of the vorticity and thermal wind as a measure of the rate of development at the 1000-mb level. Some objections were raised later to this approach by Petterssen (1955). However, it is clear that the amount of divergence is proportional to the rate of production of vorticity in an isentropic surface from the relation

$$\frac{dQ}{dt} = \frac{\partial Q}{\partial t} + \vec{V}\cdot\nabla Q = -DQ \tag{4}$$

where Q stands for the absolute vorticity.

The local rate of change of thickness $\bar{h}$ or the relative vorticity tendency of the thermal wind is expressed as the sum of advective, dynamic and diabatic processes as

$$\frac{\partial \bar{h}}{\partial t} = R \int_{p}^{p_o} [-(u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y}) + \frac{dp}{dt} (\frac{r}{g\rho} - \frac{\partial T}{\partial p}) + \frac{1}{c_p} \frac{dW}{dt}] \, d \log p \quad (5)$$

where the symbols are defined on page 144.

It is noteworthy that the strength of the divergence in an upper wave is usually proportional to its amplitude, and possibly to such dynamic properties as the generalized vertical velocity $\frac{dp}{dt}$, which increases as the wave amplifies. Hence, increase in the amplitude of the upper level contours, and divergent flow aloft will evidently produce a large amount of vorticity, a major mechanism in the development of sea-level cyclones.

CHAPTER II

RECENT INVESTIGATIONS

2.1 The Strength of Upper Air Motions

A number of studies of the behaviour and the structure of extratropical cyclones were carried out with the aid of much improved upper air soundings which became available a few years before the Second World War. The superposed short-wave perturbations on the planetary-scale long waves were investigated extensively, while some laboratory simulations of atmospheric behaviour were carried out by Fultz (1952), Fultz and Kaylor (1959) and others. Significant contributions to the development of planetary wave theory were also made by Namias and Clapp (1944), Cressman (1948), Petterssen (1950), Palmen (1951), Riehl et al. (1952) and several other researchers of the Chicago school.

From a study of long waves, Cressman (1948) observed a wave development he called "discontinuous retrogression". He found that secondary cyclogenesis frequently occurs in connection with the deepening of a minor upper trough which then becomes a new major trough upstream from the former principal trough position, while the old trough moves eastward, and the wave number increases.

As discussed in an earlier section, the strong upper air motions and the role of baroclinicity in sea-level development were emphasized by V. Bjerknes and collaborators (1911). A new and

important mechanism for sea-level development was introduced by Riehl et al. (1952) with the "jet stream". Quoting from his earlier 1948 paper Riehl states (in part):

> ... the jet stream appearing in connection with a pattern of very long waves in the westerlies provides a mechanism for the initiation of cyclone development ... the jet is effective only if it is superimposed on a disturbance of the lower atmosphere. Clearly, the amount of cyclonic development to be expected, depends in large measure on this factor. Therefore, when jet stream, long wave pattern, and low tropospheric disturbance coincide in a favourable sense, ensuing cyclone developments will attain the greatest intensity.

The strengths of the upper level divergence and associated middle tropospheric vertical motion, which in general produce the surface pressure falls, are largely influenced by the baroclinicity in the westerlies. Adequate divergence and vertical motion are necessary prerequisites for cyclogenesis and the earlier stages of cyclone intensification. Reiter (1963) remarked that such an initial wave pulse has to come from potential vorticity advection along the jet stream.

Testing Scherhag's classical idea of the "diffluent upper level pattern" an important examination was made in Germany by Polster (1960) of a large number of cyclonic developments; the main results of this work were summarized by Palmén and Newton (1969). Polster observed that 73 per cent of the cases of deepening cyclones were found under diffluent upper level contours, predominantly ahead of upper troughs; the remaining cases of initial cyclogenesis were occurring beneath parallel or confluent flow aloft, and subsequent deepening was a rare event.

Petterssen (1950) noted that, on the average, the most vigorous and frequent cyclogenesis occurs along the polar-front jet

stream[1]. Shortly thereafter, Palmén (1951) was able to conclude that the sea-level mobile cyclones are found in close association with the baroclinic region of the jet stream.

2.2 The Petterssen Development Scheme

As mentioned already in section 1.5, Sutcliffe (1947, 1950) investigated the role of the thickness pattern and vorticity in the development of pressure systems. Petterssen (1950, 1955, 1956) extended the theory and elaborated, in particular, on the development mechanism.

From the study of a generalization of the vorticity theorem and comparison with a statistical analysis of the behaviour of pressure systems, Petterssen (1950), using 41 years of hemispheric data, concluded that the frictional and the diabatic thermal components of the vorticity sources and sinks in general must balance in global isentropic sheets, even though a *local balance cannot be attained.* He also pointed out that various local circulation systems were associated with mountain ranges, inland water bodies, and other topographic features.

To explain the statistical results, Petterssen derived the following form of the vorticity equation for an isentropic surface. With some simplifications and the symbols defined on page 144, he obtained the relation

$$\nabla \cdot (Q\vec{V}) = F - \frac{\partial Q}{\partial \theta} \cdot \frac{d\theta}{dt} \qquad (6)$$

[1]More details will be given in section 2.3.

This equation shows that the vorticity budget of the atmosphere is a measure of the vorticity production or, more specifically, the vorticity export along the isentropic surfaces expressed in terms of the vorticity production of the frictional forces and the local intensity of the heat source. The vorticity production by the frictional forces and heat sinks in the lee of mountain barriers (one of the main purposes of this study) will be discussed further in Chapters 4 and 5.

Following the derivations of Sutcliffe (with Forsdyke; 1950) and Petterssen (1955, 1956), the so-called development equation may be written as

$$\frac{dQ_o}{dt} \simeq \frac{\partial Q_o}{\partial t} = A_Q - \frac{R}{f} \nabla^2 \left(\frac{g}{R} A_T + S + H\right) \qquad (7)$$

where the symbols are again defined on page 144.

From equation (7) it will be seen that *the sea-level vorticity, as a measure of pressure changes at a certain place, is primarily determined by* A_Q *the vorticity advection aloft, and by the Laplacian of the thermal components.* Furthermore, the Laplacian terms represent a local change of the relative vorticity of the thermal wind with time, in other words, *the terms reflect the importance of fronts in low level development.*

Petterssen made the subsequent simplification that the Laplacian of the thermal components may be neglected in the initial stages of cyclone development when thermal advection is relatively small. However, the Laplacian term is likely to be very effective in the formation of maritime cyclones, as discussed by Petterssen et al. (1962). Neglecting

the Laplacian term[1], the development equation becomes

$$\frac{\partial Q_o}{\partial t} \simeq A_Q \simeq -V\frac{\partial Q}{\partial s} \qquad (8)$$

where V and s measure, respectively, the horizontal wind and distance along the contours at the level of non-divergence. In addition, the vorticity advection is positive when the wind blows from high to low values of the vorticity. Now it is evident from this simplified equation that the vorticity advection at the level of non-divergence may contribute significantly to the sea-level development of cyclones.

Since $\frac{dQ}{dt} \simeq -DQ$ by equation (4), it is seen that

$$D \simeq A_Q / Q \qquad (9)$$

It is clear that the divergence in the upper troposphere, a classical measure of the development of cyclones, may be expressed in terms of vorticity relationships. The original papers should be consulted for more details.

Petterssen (1955) used equation (7) to formulate a hypothesis, now known as an important synoptic forecasting rule, which states that:

> Cyclone development at sea level occurs when and where an area of appreciable positive vorticity advection in the middle and upper troposphere becomes superimposed upon a slowly moving or quasi-stationary frontal zone at sea-level.

This rule was tested empirically on a large number of cases of cyclogenesis in North America by Petterssen et al. (1955) and later by many authors and forecasters. The tests generally confirmed that *almost all*

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[1]This term may be frequently neglected in routine synoptic work, particularly in the early stages of cyclogenesis.

cyclogenesis at sea-level occurs in advance of an upper trough, an area of divergence and positive vorticity advection in the upper troposphere. However, "vergences"[1] of considerable strength seem to be necessary in the lower and upper troposphere for favourable cyclogenesis. In the broad sense, Petterssen's rule and hypothesis agree qualitatively well with the classical development theories, and have withstood the test of the time in practical weather forecasting.

2.3 Frequencies of Cyclogenesis and Cyclone Tracks

Since the 1897 work of Bigelow (see Petterssen, 1956) and Shaw (1906) there have been many studies of the frequencies of cyclones and their tracks. Some of the most striking results were deduced by Petterssen (1950) from an analysis of vast amounts of data.

Petterssen showed that maxima of cyclogenesis occur in close association with certain geographical features, such as mountain ranges, coast-lines, bays and inland bodies of water. More particularly, he found that the primary maximum of cyclogenesis frequency occurs between the latitudes 45-50° N in the summer and between 35-40° N in winter, while a secondary maximum is located near 63° N in all seasons where disturbances may be associated with arctic frontal systems. Petterssen also noted that the primary zone of cyclogenesis occurs in a régime characterized by strong vertical shear of the zonal wind, in other words, an upper front associated with a jet maximum superimposed upon low level frontal zones with pronounced vergences and vertical motions.

Though some authors examined 500-mb level winds, Klein (1958)

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[1] Reiter seems to have been the first to use the expression "vergences" and vergence fields as collective terms for divergence and convergence. See Reiter (1963) page 328.

carried out a study at the 700-mb level. He concluded that a principal axis of cyclogenesis lies very close to the latitude of the 700-mb jet axis during most of the year in the northern hemisphere. Furthermore, the maximum cyclone occurrence coincides nearly with the axis of maximum cyclonic geostrophic vorticity.

As indicated in the general discussion in section 2.1, the results of careful studies strongly indicate a tendency of cyclones to be formed most readily in the vicinity of the jet stream where meridional and vertical gradients of the meteorological elements are pronounced. The findings also agree well with dynamic instability theories of cyclogenesis and the classical polar-front theory.

It is also of interest to examine Klein's numerical work on cyclogenesis for some indications of an annual cycle in the behaviour of pressure systems. There is a minor maximum occurrence of cyclogenesis in April and another in October-November, while a single minimum appears in September. Moreover, there are, in the northern hemisphere, on the average at least three cyclones formed for every two anticyclones, the life span of cyclones being approximately four days.

Though mean charts of cyclone tracks and frequencies of cyclogenesis are available, they are usually not of much use in the study of individual local cyclones, since the initial stages of cyclone development and motion in the lee of mountain ranges are often erratic and ill-defined.

2.4 Lee Cyclogenesis

2.4.1 Introduction

The lee of the Rocky Mountains has long been known as an important region of cyclogenesis. Other regions favourable to lee cyclogenesis are found wherever strong upper air flows cross major mountain ranges. Such flows exist on a grand scale over the Himalayas, the Alps, and the Andes, and on a lesser scale, over Scandinavia, Korea, the Pyrenees, etc.

Shaw as early as 1906 suggested that air flow could be stretched isentropically during descent in the lee of a mountain and acquire thereby some degree of instability. This is clearly suggestive of a possible mechanism for lee cyclogenesis, but Shaw did not seem to have considered its implications for development.

V. Bjerknes and collaborators (1911) also demonstrated the occurrence of stretching and convergence on the lee slope and described the production and characteristics of eddy motion in the following manner:

> A motion going on without eddies is kinematically possible, but dynamically unstable, and has therefore no chance of persisting even if it be produced for a moment ... and on the leeward side of a mountain the eddies being the most frequent .... Eddies having a vertical axis may be formed in the same way. This kind of eddy will be very frequent in the atmosphere .... The eddies can exist on every scale, down to the smallest, which must be considered as local disturbances.

Bjerknes' suggestion about the eddies being due to the instability of the motion could have provided a basis for the future development of this topic. It is interesting to note that in the analysis of sea-level pressure fields, V. Bjerknes et al. (1910) did not draw the isobars through the mountains, but indicated the discontinuities in the pressure

field by means of dotted lines.

Discussing the synoptic process of low-level development with a two-kilometer high mountain range, Shaw (1920) remarks that cyclones die when they pass over a large mountain range, and that, undoubtedly, a mountain range would cut off the bottom section of the cyclone, unless it could rapidly extend itself to the surface of the lee; but the dynamical conditions do extend to great heights, cyclones do cross mountain ranges and re-establish themselves on the lee, though many perish. In the case of the Rocky Mountains it is of course well known that fast moving cyclones and cold core lows in particular can move across the barrier and intensify on the lee side.

2.4.2 Some Earlier Synoptic Studies

Considering lee cyclogenesis in the light of new aerological data, J. Bjerknes and Holmboe (1944) concluded that the divergence fields could be superimposed on baroclinic waves in a westerly current over mountains. In this situation the vertical motion term in the Margules-Bjerknes tendency equation (1) can not be omitted on the sloping surface of mountain barriers. Bjerknes and Holmboe investigated in particular the lee development processes in the Rocky Mountains and found that, when an upper baroclinic wave with a divergence field moves eastward while a low level trough exists over the mountain range, the strong descending motion on the lee slopes induces some horizontal convergence in the lower levels of the air column, but usually not enough to prevent a fall of pressure at the lee surface[1]. If the

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[1] See Palmén and Newton (1969) page 347.

divergence field moving in over the lee side and the orographic downward motion combine, there will be a strong pressure fall. Sudden intensification depends thus on the phase lag of the upper trough. This seems to be a plausible explanation of the fundamental synoptic process of the lee cyclogenesis.

Another study of the atmospheric pressure changes on lee slopes was made by Fleagle (1947, 1948) who considered the orographic vertical motion field. Large vertical motions were found with a wave over the lee slope, but weak upward motions were associated with the earlier stage of cyclogenesis below the 3-kilometer level over a large area of the lee, and the most rapid sea-level pressure fall occurred some 800 kilometers east of the pressure trough in the case of the Colorado lee.

Hess and Wagner (1948) carried out an analysis of potential temperature fields associated with lee cyclogenesis. They observed that the configuration of the adiabats agrees strikingly well with the theoretical profiles of mountain waves of Queney (1948), which require vertical shrinking of air columns over a mountain ridge and strong stretching in the lee some distance downstream from the ridge. They also showed that, *as an air column descends the lee slope, the amount of cyclonic vorticity generated is directly proportional to the change that takes place in the column depth.* This view also agrees qualitatively with later studies of Petterssen (1950) and Bolin (1950). From the synoptic situations for lee development, the twofold effect of a Pacific low entering the West Coast was described in the paper of Hess and Wagner as follows: First, the low increases the flow over the mountains with consequent intensification of the standing lee

trough and second, the low then passes over the mountains and produces additional lee pressure falls. A number of fast moving cyclones were observed over the mountains, apparently not associated with any fronts, but they became frontal cyclones as soon as they moved out of the source region.

2.4.3 Petterssen's Application to Lee Developments

As mentioned earlier, Petterssen (1950) enunciated the important result, based on the statistical analysis of 39,691 cases of cyclogenesis in the northern hemisphere, that the principal maxima of cyclogenesis frequency occur in the lee of major mountain ranges crossed by strong westerlies. The maxima are maintained in all seasons. As it will be seen in Fig. 2, the most pronounced maximum in North America lies in the lee of the Alberta Rockies while the others are located in Colorado, and in the Sierra Nevada where thermal lows frequently occur as well. There is a distinct trough-like extension of the isolines of frequency towards the northern part of the lee of the Canadian Rocky Mountains in the summer and, to a lesser extent, also in winter.

Petterssen (1950, 1956) examines the effect of a mountain barrier on the flow pattern on an isentropic surface, and shows that

$$\frac{\partial Q}{\partial t} = - \nabla_3 \cdot (Q\vec{V}) + Q \frac{\partial w_n}{\partial n} \qquad (10)$$

where the symbols are defined on page 144. It is seen, therefore, that *the absolute vorticity production by the wind* (3-dimensional) in a unit volume on an isentropic surface *is largely determined by the intensity of the vorticity source, i.e., by the stretching and shrinking*

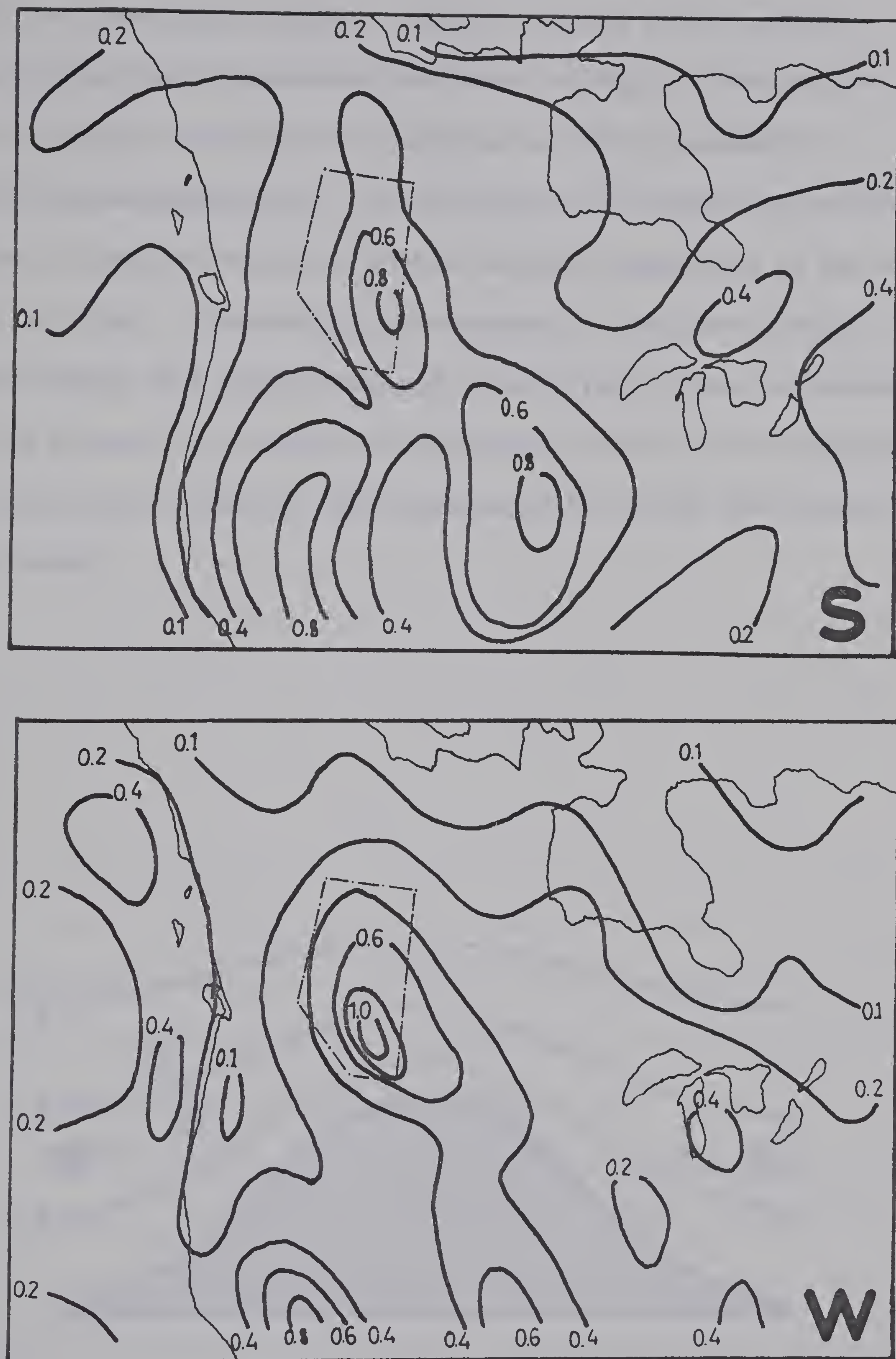

Fig. 2. Percentage frequency of occurrence of cyclogenesis in squares of 100,000 km^2 in summer (upper diagram) and winter (lower diagram). After Petterssen (1950).

normal to the isentropic surface. The air columns are in general stretched in the lee of mountains, as shown in Fig. 3. Substantial amounts of positive vorticity are produced as well by adiabatic warming of the descending air. In other words, the amount of vorticity production is largely determined by the vertical stretching of the air over the lee slope. Interestingly, according to Schallert (1962), Petterssen thinks that such stretching is occurring rather infrequently, despite the presence of strong westerly flows, and that lee cyclogenesis is to be associated primarily with peculiarities in the flow across mountain ranges.

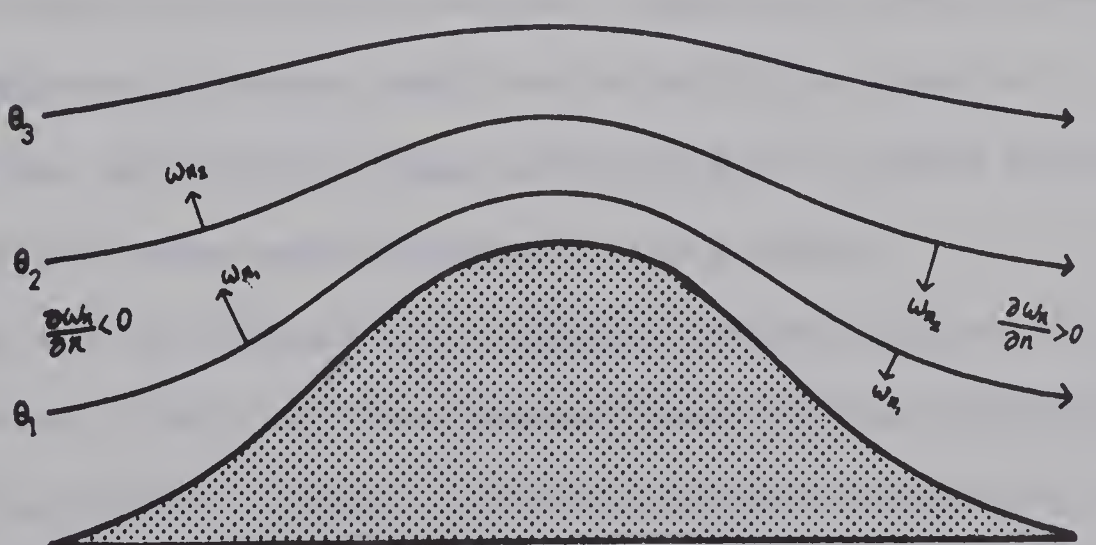

Fig. 3. A schematic profile of vertical shrinking and stretching associated with an airflow over a mountain.

2.4.4 Quasi-Stationary Baroclinic Effects

In many cases, shallow baroclinic layers may exist between air masses for a long time in the lee of the Rocky Mountains without being necessarily cyclogenetically active. This is stressed by Hage (1957, also direct communication) and by Newton (see Reiter; 1963). The baroclinicity of the basic current is claimed as the necessary requirement for both the initial deepening of a cyclone and for the maintenance of its intensity. Again, Petterssen (1955) states that the baroclinicity is a necessary and essential, but not a sufficient condition, for cyclogenesis.

The baroclinicity in general is intensified if a jet stream with strong ascending motions is superimposed upon a low level front over flat terrain, as well as over a baroclinic region on a lee slope. It is also observed that the locale of cyclogenesis is closely associated with a polar-front, as mentioned earlier, even in a region where the front is relatively shallow. Palmén and Newton (1969) found that the majority of fronts were 'dry' with little cloud or precipitation, and did not extend above the 800 or 700-mb level in the lee, though they were well-developed at the surface.

It may be inferred from classical studies such as of V. Bjerknes et al. (1911) and from mountain wave theories which will be discussed in section 2.5 that, on the lee slope of mountains, a rapid generation of low level vorticity by the incoming winds has the same role as a low level front. Thus it is quite feasible that *the Canadian Rocky Mountains, with a height of three kilometers (say to 700-mb), may create as much shearing motion as a low level front.*

2.4.5 Elaborations of the Chicago School and Some Other Results

Many studies on lee development centered on Petterssen's development hypothesis were carried out by Chicago researchers, including Newton (1956), Hage (1957, 1961), McClain (1960) and Schallert (1962).

Since the positive geostrophic vorticity advection aloft is associated with the upper wave, Hage considered the role of the upper cold low as a tool for forecasting lee development. His test of the Petterssen skeleton hypothesis involved the study of 102 upper cold lows associated with intense summer cyclogenesis in the lee of the Canadian Rockies. He discovered that "the forward march" of upper level vorticity systems is an indication of the place and time for forecasting lee development, and that the main effect of the mountain range is the role of a quasi-stationary frontal zone.

Newton, McClain and Schallert provided further support for the hypothesis, and in a broad sense they came to the same conclusions. By including orographic effects on the lee slope of the Rocky Mountains, the vertical motion field could be explored, and shown to have a pronounced effect on lee development. It is generally agreed that the 'orographic' influence is an important mechanism in the initial phases of intense lee cyclogenesis and short-lived local disturbances, and that the vorticity advection aloft is an important factor during the period of intense cyclone development.

In general, the movement of an upper level divergence field across a broad mountain range produces a large amount of low level convergence in the lee[1]. A pressure fall is likely to be maintained

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[1]Examples will be given in Chapter 4.

as long as the upper level divergence exceeds the low level convergence.

In the lee of the Alps, a semicircular barrier with a mean elevation of 2.5-3 km, Radinović (1965a, 1965b) using a somewhat different approach, observed that the maximum frequency of cyclogenesis is in the Gulf of Genoa, but not over the warmest waters. The occurrence of cyclogenesis is typically connected with a cold air outbreak towards the warm western Mediterranean basin. He also noticed the processes of low level development from the deformation of the thermal and flow patterns by the Alps. The passage of the thickness trough of the layer 1000-500 mb causes an increase of the baroclinicity in the lower layer, and thus in an increase of positive vorticity. The study could be considered a test of mechanisms proposed by Bjerknes, Exner and Sutcliffe.

The orographic development is also strongly influenced by the general character of the large scale flow pattern and as well by the geographic location. Bolin (1950) and Reiter (1963) showed that the large mountain ranges of the Himalayas and the Rockies induce each a quasi-stationary long-wave trough -- one at 130° E and another at 80° W. If a broad belt of westerlies flows over a meridionally extensive range of mountains, a long north to south trough can be found in the lee of the massif, frequently containing two cyclones. Such a case will be shown in Chapter 4.

From a comparison of cyclone development over the North American continent and the North Atlantic Ocean, Petterssen et al. (1962) concluded that the lee development occurs normally when a pre-existing upper cold trough with strong vorticity advection on its forward side approaches a low-level baroclinic zone, but with a *significant*

separation between the upper- and low-level systems during the early stage. By contrast, maritime cyclogenesis may begin under a straight upper current *without* appreciable vorticity advection in a strongly baroclinic region. Furthermore, the *thermal advection* in the low levels provide the major contribution throughout the development, and the *separation distance remains unchanged for maritime cyclones.*

Petterssen et al. (1962) did not succeed in constructing a satisfactory lee cyclone model which could account for the observed weather patterns, due to great irregularities and synoptic differences in the structure of the cyclones. This suggests that the application of the "classical" Norwegian model (and the associated frontal weather pattern) to lee development is likely to prove unsatisfactory.

Fawcett and Saylor (1965) studied the changes of weather patterns accompanying Colorado cyclones. By using Schallert's (1962) method for the selection of cases, Fawcett and Saylor showed that the weather increased in severity some 12 hours after cyclogenesis with the eastward movement of the system, noting that, on the average, the air was relatively dry during the first 12 hours. The most severe weather is likely to occur some 24 hours after cyclogenesis in the most intense cyclones. So far, no similar investigation has been carried out on Alberta cyclones.

Though not directly related to the present study, it is of interest to recall one of the conclusions reached in the course of a hail study by Longley and Thompson (1965). They suggested that afternoon hail is most likely to occur with southwest winds aloft, in situations when there is an upper trough over British Columbia in the morning, and a low-level warm tongue ahead of an advancing cold front

in Central Alberta. In a general sense, the mechanism for such mesoscale developments is similar to what is required for the lee development of cyclones, especially in the initial stages of cyclogenesis.

2.5 Mountain Waves

It has long been known that air flow over a mountain is usually much more disturbed than flow over a flat terrain. This was clearly understood by Shaw (1906) and V. Bjerknes et al. (1911). Extensive observations have shown that gravity waves can occur to the lee of any mountain barrier in the world. The Canadian Rocky Mountains[1], a long as well as a broad barrier, are obviously effective on every scale in disturbing the air flow.

Küttner (1939) has drawn attention to an analogy between lee eddies and föhn waves, and to the Kelvin-Rayleigh ideas concerning lee phenomena in a water stream. Such dynamical effects as increases and decreases of pressure about an obstacle could be demonstrated simply by water flowing in a channel, in accordance with Bernoulli's theorem, when the flow velocity in a fluid of depth h exceeds the critical velocity[2], $\sqrt{gh}$. If the fluid is stably stratified, the flow will give rise to wave motion in the lee. It is obvious that well-developed wave motion characteristically appears in a stable atmosphere. If the

[1] The Alberta Range has a half-width of 80-120 km, and smoothed peak heights of 2.5-2.7 km.

[2] Long waves may be generated when the fluid velocity is smaller than the critical velocity. See Queney (1948).

obstacle is a mountain range and if the flow is at right angles to it, one may expect to have shear motion and a pressure trough in the lee of the barrier, in accordance with established classical ideas.

Fig. 4, which has been adapted from Lyra (1943, and also Queney et al., 1960), shows the air flow profiles produced by a simple, weir-type obstacle. Lyra finds that a wave of maximum amplitude appears at some distance downwind from the obstacle, but the amplitude has a tendency of damping out with the propagation downstream. He also shows that the pressure disturbances in the lee side of the obstacle undergo quasi-periodic oscillations. These results, of course, cannot be applied directly to the orographic problem, due to the gross differences in scale and obstacle profile.

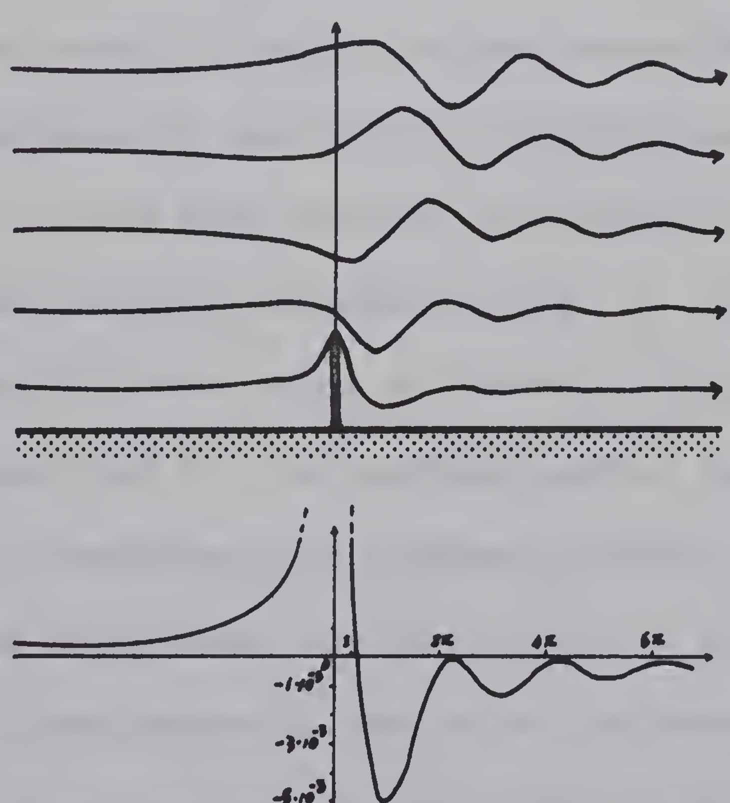

Fig. 4. Vertical stretching in the lee and relative pressure variations on the ground in the case of a small obstacle. Adapted from Lyra (1943).

Queney (1948) and Scorer (1948) made theoretical investigations on a smooth, bell-shaped mountain and found that the formation of lee waves depended mainly on the stability $(\frac{g}{\theta}\frac{\partial\theta}{\partial z})$ and the winds. With certain stratifications and the assumption of frictionless, initially undisturbed flow of 10 m/sec over a mountain range of 1 km elevation, and 10 km half-width, Queney calculated values of ±0.7 mb for the pressure fluctuation at the ground. Alternating zones of vertical stretching and shrinking are present downstream of the narrow mountain range.

In the case of a typical, broad mountain range of 100 km half-width, i.e., of a size comparable to the Canadian Rocky Mountains, the Queney profile shows again alternating zones of vertical stretching and shrinking of the air columns in the lee, and a large amplified wave at about the 3-5 km level. A region of weak stretching appears near the ground at a distance of about 400 km from the mountain crest, and a secondary layer of very weak vertical stretching is present at low-levels about 950 km downwind, as shown in Fig. 5. Moreover, it is possible that these distances would be somewhat greater if the wind increased with height and if the mountain profile were higher. Also, if strong vertical stretching with enhanced cyclonic vorticity production occurred downstream, one could expect a trough or a weak low level cyclonic development in the lee of the barrier.

The synoptic-scale Queney model, which includes the effect of the Coriolis force, shows horizontal meandering of a streamline, which may be explained by the Rossby (1940) theorem of the conservation of potential absolute vorticity. In the form,

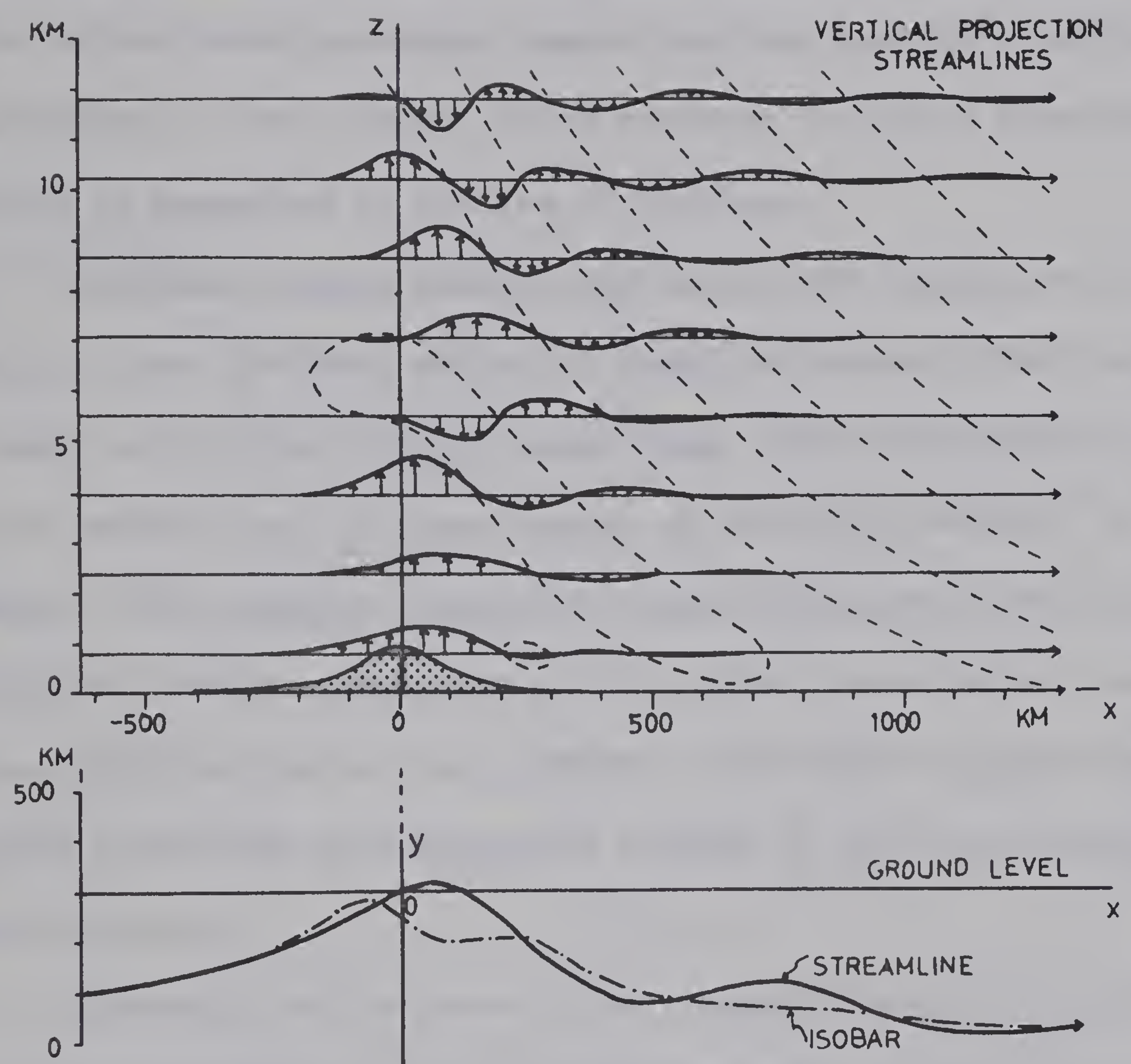

Fig. 5. Streamlines of a frictionless stratified airflow over a broad mountain range. The height of the range is 1 km, and its half-width 100 km. The undisturbed upstream speed of the current is 10 m/sec. Upper part: vertical cross-section showing pattern of streamlines. The dashed lines mark off the zones of vertical stretching and shrinking. Lower part: horizontal projection of a streamline of a particle and the pressure variations at the ground. After Queney (1948).

$$\frac{(f + q)}{h} = \text{constant} \tag{11}$$

where h is the depth of the air column. Therefore, it is clear that the flow across broad mountains always involves horizontal deflection and meandering of the current, since positive vorticity (cyclonic curvature) is generated in the lee of barriers.

The Queney models develop lee waves with troughs and ridges generally tilted upstream, while the models of Scorer (1949) produce waves which are tilted slightly downstream. The differences in tilt appear to reflect the different scales of vorticity inherent in the two models. The possible causes for these variations might stem from the different initial conditions of wind speed, temperature, mountain profiles, Coriolis force, etc. However, both models suggest that *baroclinic conditions* will generally develop in the lee of massive mountain barriers.

As pointed out by Corby (1954), certain results of Scorer's work are not applicable to large scale phenomena, owing to the neglect of the earth's rotation. Nevertheless, at least on the meso-scale shown in Fig. 6 (adapted from Scorer (1949)), it is found that the strong cyclonic vorticity produced at the 1.5 km level is well beyond the approximately 2 km half-width distance of the mountain range of 1.5 km elevation. Scorer identified this region with the level of wave clouds over the lee slope. However, in the case of large mountains it is observed that the maximum amplitude of lee waves is usually attained near the layer of high static stability, at heights of about 4-5 km.

Satellite pictures have been used in recent investigations of

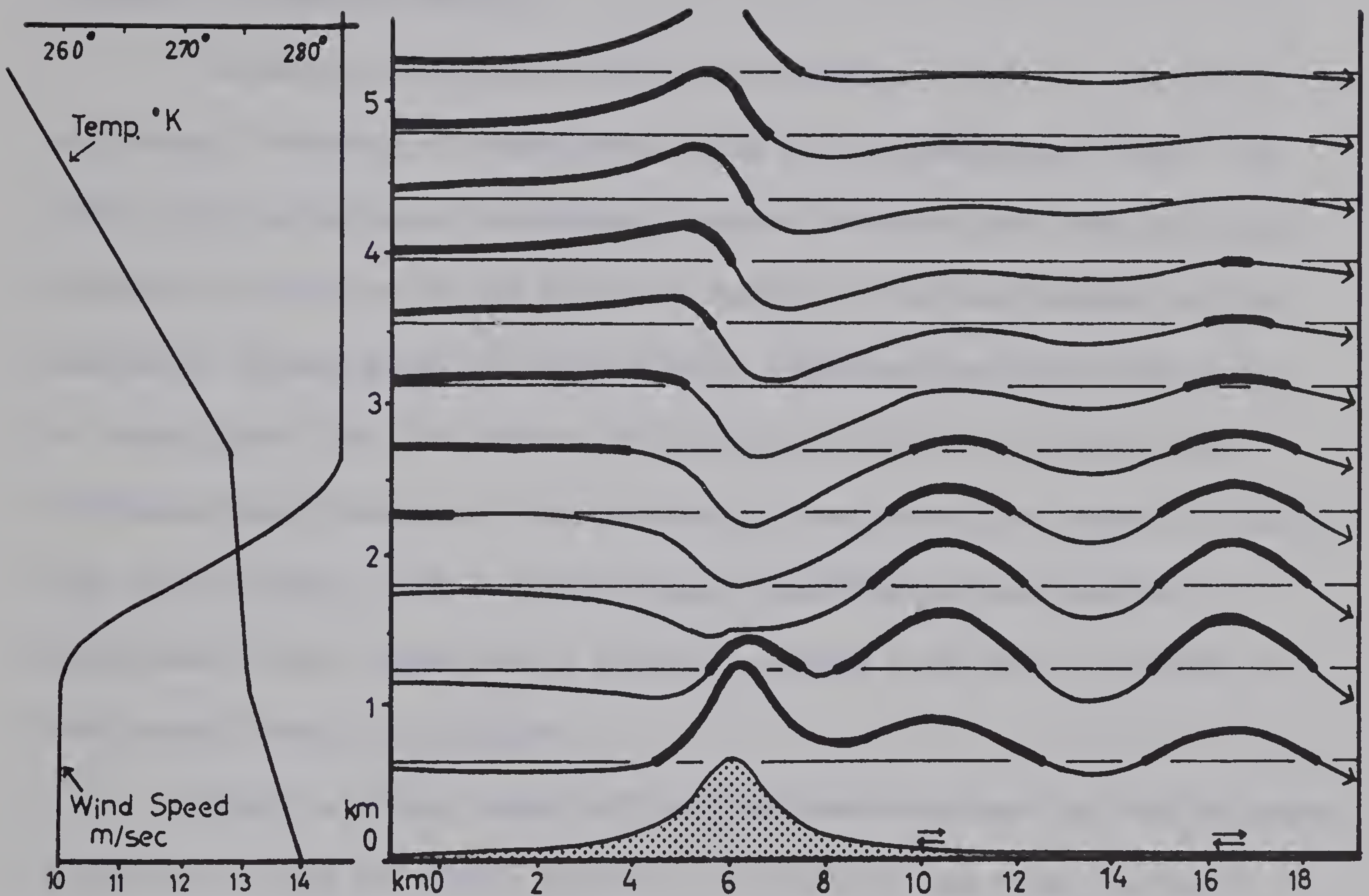

Fig. 6. A mountain wave profile with non-uniform wind speeds and temperatures. The thick streamlines represent the air above its original level. After Scorer (1949).

lee wave phenomena, e.g., by Dirks et al. (1967). It is also noteworthy that pronounced vertical stretching clearly appears in laboratory simulations of mountain waves such as were performed by Long (1959). Moreover, it is well known from wind tunnel studies that an airflow normal to an obstacle produces a larger amount of shear than flows incident at smaller angles.

Recently, Vergeiner (1971) constructed a multi-layer model employing a variety of stabilities with a two-dimensional flow. He found very satisfactory agreement between observed and computed flow patterns in the lee of the Colorado Rockies. The half-width of the mountains investigated by Vergeiner is somewhat less than what would be appropriate for the Alberta Rockies, and the induced waves are correspondingly shorter. Interestingly, lee waves are likely to occur over flat ground, with somewhat longer wave-length and smaller amplitude, rather than over a sloping surface when the air stream at the lowest levels is blocked.

While a great number of studies have been made on the Colorado Rockies and also for other regions, little work has been done with typical profiles of the Canadian Rocky Mountains.

Wallington (1955) made a synoptic study of lee waves and concluded that the amplitude reached a maximum about 900 km ahead of a surface warm front in the lee of a hill 400 meters high. From a study by Holmboe and Klieforth of synoptic situations producing lee waves in the Sierra Nevada, Queney et al. (1960) noted that strong lee waves are associated with an upper trough along the Pacific coast, with westerly flow, and a cold front or occluded front approaching the region from the northwest. Furthermore, *strong lee waves occur*

occasionally with a jet stream lying across the mountains. Thus it seems plausible that situations conducive to lee waves are also favourable for lee cyclogenesis.

In any case, these diverse theoretical and empirical results lend support to the generally accepted view that, *for the formation of well-developed lee waves, the air flow must cross the mountain ridge more or less perpendicularly with considerable speed and depth, and with some persistency.* Moreover, the orientation, the shape, and the height of mountain ranges together with the conditions of the atmospheric flow determine the wavelength of the atmospheric disturbance, and thus quite possibly the frequency and nature of lee cyclogenesis.

Almost all lee wave models, though largely products of fluid-dynamical theory, are in good agreement with lee wave observations, and *most show a zone of enhanced vertical stretching some distance beyond the half-width of the mountain profile. If this zone were to be identified with a region of cyclogenesis associated with synoptic-scale barrier waves, the centre of maximum activity should be located well downstream in the lee of the massif.*

2.6 Energy Transformations in Cyclones and Numerical Approaches

Though the present dynamical theories are of little practical use in routine work, they are helpful in understanding some of the behaviour of cyclones. Cyclogenesis has long been, and still is, a major problem in the development of numerical forecasting, despite the great advances made since the days of L. F. Richardson.

When considering the problem in terms of baroclinic air flow, Thompson (1961) suggested that cyclogenesis may be regarded as a type

of hydrodynamical instability (amenable to treatment by the perturbation methods for amplifying small baroclinic waves) and as a process of energy conversion, since the accumulated internal and potential energy of a cyclone is transformed into kinetic energy.

When constructing one of the early numerical models embodying exact stability criteria, and using the development principle of J. Bjerknes (1937), Charney (1947) assumed the actual flow to be a small perturbation superimposed on the westerly flow in an adiabatic, compressible atmosphere. Linearized perturbation equations of motion were also used in the model. He found that "westerlies of middle latitudes are a seat of constant dynamic instability" with a maximum of low-level horizontal convergence occurring between the upper trough and preceding ridge.

Following this study, a number of numerical investigations have focussed on the solution of the problem of cyclogenesis, for example Charney and Eliassen (1949). The scope of those studies is beyond the purpose of the present study, but the details are available in the original publications.

In the recent studies of Kung (1966) and of Palmén and Newton (1969) it is stressed that kinetic energy is generated mainly within the disturbances, and that this requires sinking of cold air and rising of warm air. A large amount of kinetic energy is generated in mature cyclones, but the source of this energy is mostly the potential energy of the environment. Moreover, energy transformation occurs in the upper and lower troposphere and the kinetic energy is eventually dissipated by frictional processes.

It is of interest to note that an important source of kinetic

energy is found in the regions of horizontal divergence, as pointed out by Starr (1948). Sechrist and Dutton (1970) also show that the principal sources of kinetic energy are located in a baroclinic region associated with a jet maximum in the west of a trough, and the southeast region of the trough in the vicinity of the warm sector of the cyclone. The maintenance and intensity of the general circulation depends upon the balance between generation and dissipation of the kinetic energy, a requirement in good agreement with the 9-layer general circulation model of Smagorinsky et al. (1965). In addition, the idea that principal sources of kinetic energy are associated with jet streams agrees also with a recent study by Petterssen and Smebye (1971).

An early discovery of J. Bjerknes and Solberg (1922) and emphasized anew by Palmén and Newton (1969), it is now well-known that cyclones are an essential part of the general circulation of the atmosphere. Cyclones are effecting a large part of the necessary meridional and vertical exchange of angular momentum and heat. Such classical ideas now permit the construction of global general circulation models such as that of Phillips (1956), and of many others.

Much effort has also been recently expended in the use of the primitive equations of motion in certain prediction models, e.g., Smagorinsky et al. (1965) and Shuman and Hovermale (1968), superceding in part the systems of linearized differential equations which had been used in applications of Rossby's (1940) vorticity theorem, and the like.

Furthermore, orographic effects have been incorporated into many numerical prediction models, such as those of Sawyer (1959), Cressman (1960), Graystone (1962), Fawcett (1969), Miyakoda et al. (1969), and Holloway and Manabe (1971) In particular, Fawcett showed

that better forecasts of cyclogenesis in Western Canada could be obtained by the introduction of more realistic mountain topographies, rather than by the use of heavily-smoothed barrier profiles. This indicates that in a numerical weather prediction model the incorporation of better, not overly-smoothed topographies is necessary if forecast errors are to be reduced.

2.7 Summary

According to Margules (1904), V. Bjerknes et al. (1911), Dines (1919), Scherhag (1934, 1937) and J. Bjerknes (1937, 1944), it seems that upper level divergence and low level convergence are necessary for sea-level cyclogenesis. In addition, if a strong baroclinic region with a jet stream is present in the incipient synoptic pattern, the development is much more probable, as shown by Petterssen (1950, 1955), Riehl et al. (1952) and Reiter (1963). Moreover, a hyper-baroclinic zone usually appears in the lee of any broad mountain system even with a quasi-uniform air flow over the barrier, in accordance with the studies by V. Bjerknes et al. and the mountain wave theories of Lyra (1943), Queney (1948), Scorer (1949) and others. From Petterssen's development equation, it seems that upper vorticity systems may largely determine low level cyclone intensification in the lee of the Canadian Rockies.

It is likely that, by way of a generally operative process, vorticity is produced by vertical stretching of the descending winds in the low levels of lee slopes. This generates lee troughs which remain stationary until they are overtaken by a divergent region ahead of an upper trough. Thus, the lee of a massive mountain system is a favoured region for the generation of cyclonic disturbances.

Many studies suggest that orographically induced motions appear to be very important as the mechanism producing the vorticity required for lee disturbances, as emphasized by Petterssen (1950, 1956). According to the Sutcliffe and Godart (1942) equation, it appears plausible that, in general, but especially in the case of massive mountain ranges, orographic vertical motions might affect the vergence fields significantly, when the air flows for a comparatively long time over the barrier. This will be discussed in Chapter 5.

Queney (1948) and Bolin (1950) considered the case of a relatively strong current striking a mountain range at right angles, and their results indicate that the shape of mountains is important in the development of lee disturbances. Hage (1957) formulated that the favoured conditions for intense summer lee cyclogenesis are present when upper cold lows drift inland from the west coast of the Continent. In the case of the Canadian Rocky Mountains it appears that a stable, strong, divergent flow of considerable depth and roughly perpendicular to the Continental Divide is required for lee cyclogenesis. This mechanism will be examined in more detail in section 3.8.

CHAPTER III

THE SAMPLING OF CASES AND TESTS OF DEVELOPMENT THEORIES

3.1 Introduction

Much effort was expended by several Chicago researchers investigating the mechanism of the *sudden lee intensification* of major cyclones (discussed in the section 2.4) which requires positive vorticity advection aloft, preferably ahead of, and associated with, upper troughs and cold lows. However, weak cyclonic systems, variously referred to as "local disturbances", orographic lows, heat lows, and the like, have received comparatively little attention, and not many investigations aimed at the better understanding of the fundamental problem of the *initial formation* of lee cyclones can be found in the literature. Of course, some important if not basic mechanisms of cyclone behaviour remain as yet to be fully explained.

It is the purpose of the present study to investigate whether cyclones generated in the lee of the Canadian Rocky Mountains developed in accordance with the aforementioned development theories. A cyclone was considered to be a lee development if cyclogenesis occurred in the lee of the Rocky Mountains anywhere between Montana and the mouth of the Mackenzie River. The total area covered by the study was very large. However, the main concentration was devoted to the area bounded by the parallels of latitude 40° and 70° N, and by the meridians 80° and 160° W as shown in Fig. 8. The cyclones examined were so diversified

in size, behaviour and complexity, that it was necessary to classify them according to certain specified criteria. The several types of lee development were, on the average, traced back four days prior to a cyclogenesis, and followed forward for three days after the cyclogenesis had occurred in the lee.

3.2 The Topography of the Canadian Rocky Mountains

3.2.1 The General Topography

The Fig. 7 shows the topography and general configuration of the Canadian Rocky Mountains. The main block of the Mackenzie Mountains stands somewhat apart and some two degrees of longitude to the east of the principal ridge of the Rocky Mountains. It will be seen that the Canadian Rockies run roughly northwest to southeast, an orientation which places them athwart the prevailing westerly to southwesterly upper air circulation.

The Canadian Rockies consist of "three principal mountain systems" separated by major passes, as shown in smoothed form in Fig. 8. *From south to north these principal systems are the Southwestern Alberta Range, the Northern British Columbia Range, and the Mackenzie Mountains.* In the same order, the three ranges are oriented as follows: 140°-320°; 150°-330°; and 135°-315°. Three major passes cross the mountains, respectively, near the Alberta-Montana border, west of Dawson Creek, B.C., and in the vicinity of Watson Lake, Y.T.

The Southwestern Alberta Range of the Rockies, which includes the National Parks of Jasper and Banff, is flanked on the west by the

Columbia Mountains which comprise the Selkirk and Cariboo Mountains[1]. The highest peaks are over 3.5 km, the most prominent being Mounts Assiniboine, Temple, Forbes, Columbia, Alberta and Robson. In addition, there are many peaks higher than 3.0 km in this range.

The second principal mountain system, the Northern B.C. Range, is a broad complex connected in the west to the Coast Mountains by the Omineca and the Cassiar Mountains. The main peaks are Mt. Lloyd George and Mt. Churchill, both over 3.0 km high. The Mackenzie Mountains, the smallest of the principal ranges, consist of the Pelly, Selwyn, Ogilvie and Richardson Mountains. The highest peaks are Dome Mountain, Sir James McBrien and Mt. Keel, all over 2.7 km in altitude.

The highest mountains in North America are, of course, located in Alaska, and much of the interior of Alaska comprises relatively high terrain. But the actual Continental Divide of the Arctic Ocean - Pacific Ocean watersheds is found on the Richardson Mountains, where a few peaks are some 1.5 km high.

The movements of cyclones in Alaska and the Yukon are irregular, and difficult to trace. This probably reflects the erratic distribution of the major Alaskan and Yukon ranges: most are oriented west to east, some southwest to northeast, and a few, mainly along the Divide, northwest to southeast. Thus the lee regions of most ranges are poorly defined with respect to the prevailing westerlies. For this reason, the study of lee developments in the Mackenzie Mountains was limited to the lee of the Richardson Range.

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[1] The names of some mountain ranges differ from atlas to atlas. For consistency of description, the names given in the Canadian Oxford Atlas are used here.

Topography of the Canadian Cordillera

At first glance, the general aspect of the lee of the Canadian Rockies appears to be that of a huge plain dipping gently toward east, but detailed geographical maps show that the terrain features are far from simple. In the case of each principal mountain system, the 1000-meter height contour meanders widely in the lee, and two river basins penetrate deeply into the major passes. The 500-meter contour runs roughly along a line from Fort Vermilion to Prince Albert and Brandon. The horizontal distance between the Continental Divide and the 500-meter contour varies from about 500 km to 900 km. But the lee of the Mackenzie Mountains is somewhat steeper, and a broad, flat low-land opens up rather suddenly to the east, comprising the Mackenzie Valley, Great Bear Lake and Great Slave Lake.

Furthermore, there are several blocks of hills and low mountains well to the east of the Divide, such as the Swan Hills, the Clear Hills, the Birch and Caribou Mountains, the Cameron Hills, and the Horn Mountains, all of which tend to affect the development of pressure systems in some degree.

3.2.2 Construction of a Smoothed Topography

It is clear that because the Canadian Rocky Mountains are quite complex, the elevation of a given point may not represent the surrounding area adequately. However, synoptic-scale air motions over mountain barriers seem to follow, by and large, the contours of the average, "smoothed" terrain. Hence it is desirable and necessary to construct a smoothed terrain profile in order to evaluate the vertical motions induced by the sloping terrain. This was recognized in the classical papers referred to earlier, which considered the

generation of vertical motions in developing systems. For example, V. Bjerknes et al. (1911) emphasized the necessity of the construction of special maps of idealized topography for mountain regions and pointed out the following:

> ....In this manner correct values will be found for the average intensity of the forced ascending or descending motion, while the small irregular motions up and down, which are only of local importance, will drop out. But it should be remembered that the drawing of the idealized charts has no unique solution.

For the present study, a smoothed topography of the Continental Divide and the area to the lee of the Rockies was prepared from topographical maps of scale 1:250,000. This topography, shown in Fig. 8, was obtained from an analysis of contour heights read off at every 2.5° intersection of latitude and longitude, with the aid of a 9-point one-half degree mesh size grid. The orientation of the grid was north-south along the meridian, and east-west along the lines of parallel. To obtain a representative height at a given point of intersection, the height of each of the 8 surrounding points was estimated by eye from the contour pattern of the topographic map. The average height at a grid point was then evaluated by the following expression:

$$\bar{Z} = 0.5\, Z_o + 0.5 \left(\frac{\sum_{i=1}^{8} Z_i}{8} \right) \qquad (12)$$

where Z_o is an unsmoothed height read off at the central grid point, Z_i is the height at a neighbouring grid point, i, and $\bar{Z}$ is the resulting smoothed height. This method of smoothing is used also in the numerical prediction model of Holloway and Manabe (1971).

At a few grid points near the Continental Divide the smoothing process was modified somewhat, if the location of the central point was

Fig. 8. A smoothed topography of the lee of the Canadian Cordillera. Contours are in hundreds of meters.

found to be too low compared with its surroundings, in order to avoid an overly-drastic reduction in height of dominant mountain ranges. Thus a 3-km high range could easily be smoothed down to one-half its natural height if no precautions are taken. Such drastic reductions of height lead to errors in forecasting, as shown by Fawcett (1969) and noted earlier in section 2.6.

Fig. 8 clearly shows the three major mountain systems as well as the major passes. The smoothed peaks of the Southwestern Alberta Range are some 2.5-2.7 km high, and the half-width of this mountain range varies in the lee side from 80 to 120 km. The smoothed height of the Northern B.C. Range is of the order of 2.0-2.3 km, and has a half-width of 70 to 100 km. The Mackenzie Mountains have a smoothed height of 1.5-1.8 km, and a half-width of 120 to 160 km.

It is of interest to note that a typical slope of a major mountain system such as the Southwestern Alberta Range between the Continental Divide and the half-width distance is approximately 1:40. This is a somewhat steeper slope than that of a typical cold front. However, the average slope between the Divide and the 500-meter height contour to the east is much less, ranging from 1:170 to 1:320 approximately, slopes more typical of warm frontal surfaces.

3.3 Selection of Cases

According to Schallert (1962), Colorado lows develop mainly between late October and spring. However, since significant lee cyclogenesis does occur in Western Canada in all seasons, it was necessary to study cases of development throughout the year. The principal synoptic charts used were surface and 500-mb maps of 1958,

prepared by the Edmonton Weather Office, Atmospheric Environment Service of Canada. On occasion, it was necessary also to make use of the Historical Weather Maps published by the U.S. Weather Bureau for the International Geophysical Year.

In this study, cases of lee cyclogenesis were first identified on surface charts and then, insofar as possible, traced back and forward on consecutive 6-hourly synoptic charts. In many cases intermediate 3-hourly local charts were helpful in tracing out the history of a development. If the development was associated with an upper trough, the history of the upper circulation was also recorded in order to identify certain characteristic features[1].

In order to classify lee developments objectively, the following criteria were set up and applied consistently to low-pressure systems and incipient lows appearing in the lee of the Canadian Rockies:

(i) A low or incipient low pressure centre was present if the centre was enclosed by at least one isobar in a sea-level field analyzed in terms of isobars drawn at two-millibar intervals.

(ii) The closed isobar about the low had to persist for a period of at least 24 hours on a set of consecutive charts.

A total of 146 cases satisfied the above conditions in the primary area, and were investigated in detail.

3.4 Evaluation of Intensity of Lee Cyclones

Changes in central pressure of a cyclone often do not indicate the change of circulation about the centre, and the strongest circulation usually occurs at some distance from a cyclonic centre.

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[1]Detailed procedures of finding the history of the upper wave will be described in section 3.9.

However, the circulation may be estimated by computing the intensity J of a sea-level cyclone from the intensification formula of Petterssen (1956):

$$J \equiv \nabla^2 p \simeq \frac{p_1 + p_2 + p_3 + p_4 - 4p_o}{H^2} \qquad (13)$$

where ∇^2 is the horizontal Laplacian operator, p is the sea-level pressure, H is the grid interval, and the subscripts denote the points of a standard grid for finite-difference computation as indicated in Fig. 9.

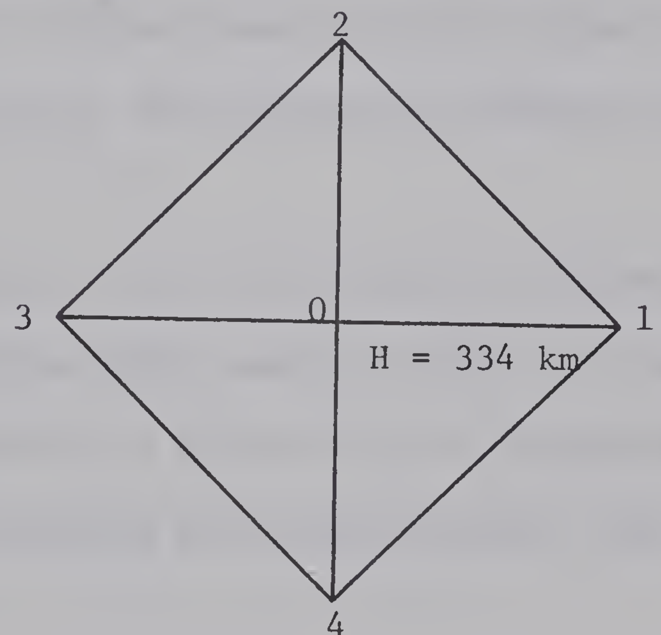


Fig. 9. A finite-difference grid with a mesh size of 334 km for the computation of the intensity of sea-level cyclones and the vorticity at the 500-mb level.

In a similar study Radinović (1965b) used a mesh size of 500 km in Europe, and Schallert (1962) suggested that a 3-degree latitude length is suitable for the purpose. Hage (1957) used a mesh size of

300 km when computing the absolute geostrophic vorticity at the centre of surface lows. The value of any quantity calculated by finite-difference methods depends to some extent on the grid length. In the case of intensity calculations, it should be also noted that the values of J may vary by as much as 20 per cent with changes of the grid orientation, when the finite-difference grid is applied to non-circular, asymmetric cyclonic systems. Moreover, the finite-difference values do not usually take full account of the size of a cyclone, but reflect only the relative intensity of the circulation in the area covered by the finite-difference grid. Thus when the strongest circulation about a cyclone centre occurs beyond the boundary of the finite-difference grid, the estimate of intensity may be at times too low.

Furthermore, in the evaluation of intensity, difficulties may arise also in regions where station reports are sparse. In such cases, the central pressure of a cyclone may be estimated from the gradient wind equation by assuming horizontal gradient flow. The expression used is

$$p_c = p_i - \frac{\rho f^2 r^2}{8} \qquad (14)$$

where p_c and p_i refer, respectively, to the pressure at the centre of the cyclone and to the known pressure at a nearby isobar, r is the horizontal distance between p_c and p_i, ρ is the density, and f the Coriolis parameter at the centre.

3.5 Classification and Grouping of Lee Cyclones According to Intensity

For the consideration of lee developments in the Alps, in 1959, Radinović (1965b) subdivided the lee cyclones into three groups, on the basis of computed intensity, in the following manner:

(i) Weak cyclones: $\nabla^2 p \leqq 20 \text{ mb}/(500 \text{ km})^2$

(ii) Moderate cyclones: $21 \text{ mb}/(500 \text{ km})^2 \leqq \nabla^2 p \leqq 40 \text{ mb}/(500 \text{ km})^2$

(iii) Strong cyclones: $\nabla^2 p \geqq 41 \text{ mb}/(500 \text{ km})^2$

However, this classification by intensity was found to be rather inadequate when applied to the typically moderate to intense lee developments in Western Canada. Instead, sample tests indicated that the grouping method of Schallert (1962) was more appropriate to conditions in the lee of the Rockies.

Schallert divided 71 cases into groups in accordance with the maximum intensities acquired during their life span. Each group was then subdivided into cyclones which either moved out of the Colorado region along well-defined trajectories, or remained there, i.e., local systems with short, erratic trajectories.

The 146 cases of the present study were grouped and classified as follows:

(i) Type A: Moderate to intense cyclones which move out of the source region (76 cases);

(ii) Type B: Moderate cyclones which remain quasi-stationary or have very short trajectories (6 cases);

(iii) Type C: Weak cyclones which move out of the source region (33 cases);

(iv) Type D: Weak cyclones which remain quasi-stationary or have very short trajectories (31 cases).

Detailed figures for each of these types of lee cyclogenesis are given in Table 1. All cyclones of Type A underwent major development. They travelled relatively long distances from the source region, and had well-defined trajectories. Cyclones of Type C also migrated away from the Divide but their behaviour was more erratic compared to that of Type A. By contrast, cyclones of Type B and D remained wholly on the lee slope throughout their life spans or at least within the area to the west of the 500-meter height contour. Type B cyclones attained moderate intensity, but they remained small and did not acquire significant circulations. Furthermore, no cyclone of Type B, C and D was directly associated with an advancing upper cold trough at any time throughout its life span. On the other hand, all the members of Type A had well-developed upper troughs and intensified rapidly.

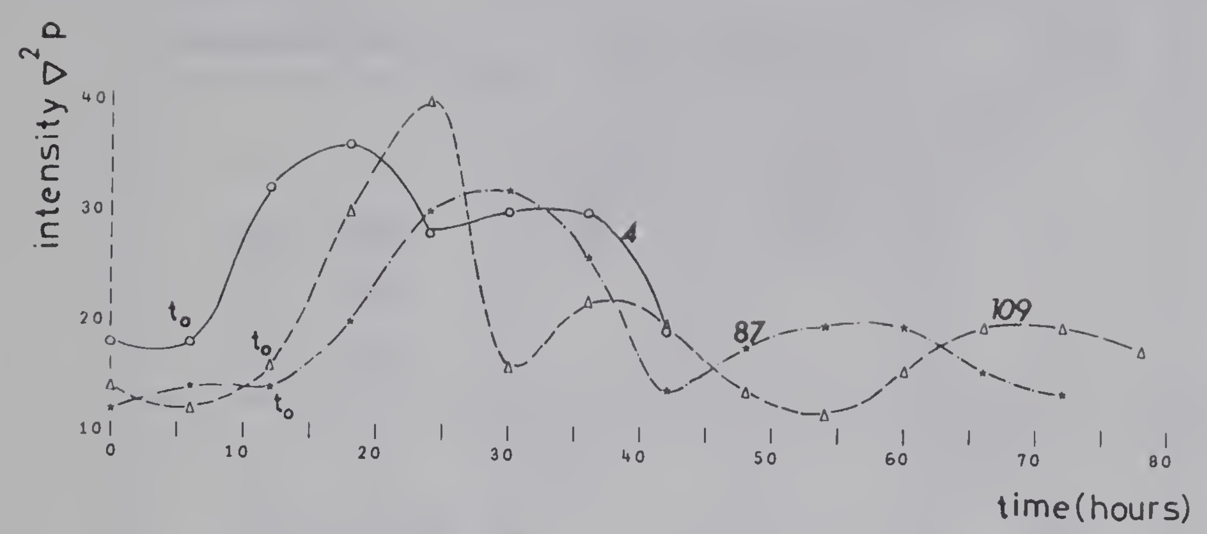
Schallert did not test Petterssen's development theory on cyclones of Type C and D. However, all types of cyclones were considered in the present study. The lesser cyclones of Type C and D, though of much lower intensity than the major storms, are nevertheless important synoptic features which contribute significant amounts of eddy flux to the budget of the general circulation. Furthermore, since all cyclones originate as small disturbances in the pressure field, the study of small synoptic-scale eddies may well lead to a better understanding of the principal mechanisms of *initial* development.

In forecasting, the time of cyclogenesis and of rapid intensification are very important. Hence, much effort was spent on investigating the mechanism of sudden intensification by researchers of the Chicago school, and others. Schallert in particular observed that some intense Colorado cyclones underwent two successive periods of

intensification. Palmén and Newton (1969) also noted this feature in the development of certain cyclones. In order to take account of such cases of two-stage development in the present study, Type A was subdivided further into Types A_1 and A_2. Type A_1 includes 48 cyclones which underwent a single intensification commencing at the onset time of intensification t_o. Type A_2 cyclones (28 cases) are those with two periods of intensification, and onset times of t_1 and t_2 respectively. Examples of these two types of intensification are shown in Fig. 10.

3.6 Frequencies of Lee Cyclogenesis in the Canadian Rockies

As mentioned earlier, all cyclones with a life span of at least one day were selected for detailed examination. Table 1 is constructed on the basis of the aforementioned classification. It is of interest to note that the developments of Type B and D constitute about 25 per cent of the total sample. The other 75 per cent of lee developments in the Canadian Rockies involve relatively long-lived systems with long trajectories toward the eastern regions of the continent. Moreover, about one half of the total cyclones acquired at least moderate intensities and were associated with upper troughs for at least part of their life spans. These percentages differ considerably with Schallert's results of Colorado cyclones. For example, he found the number of local developments in Colorado to be about 45 per cent of the total. These marked changes in percentage frequency are probably due to differences in the height and shape of the two mountain systems, and in the diverse strength and structure of the cross-barrier flows.



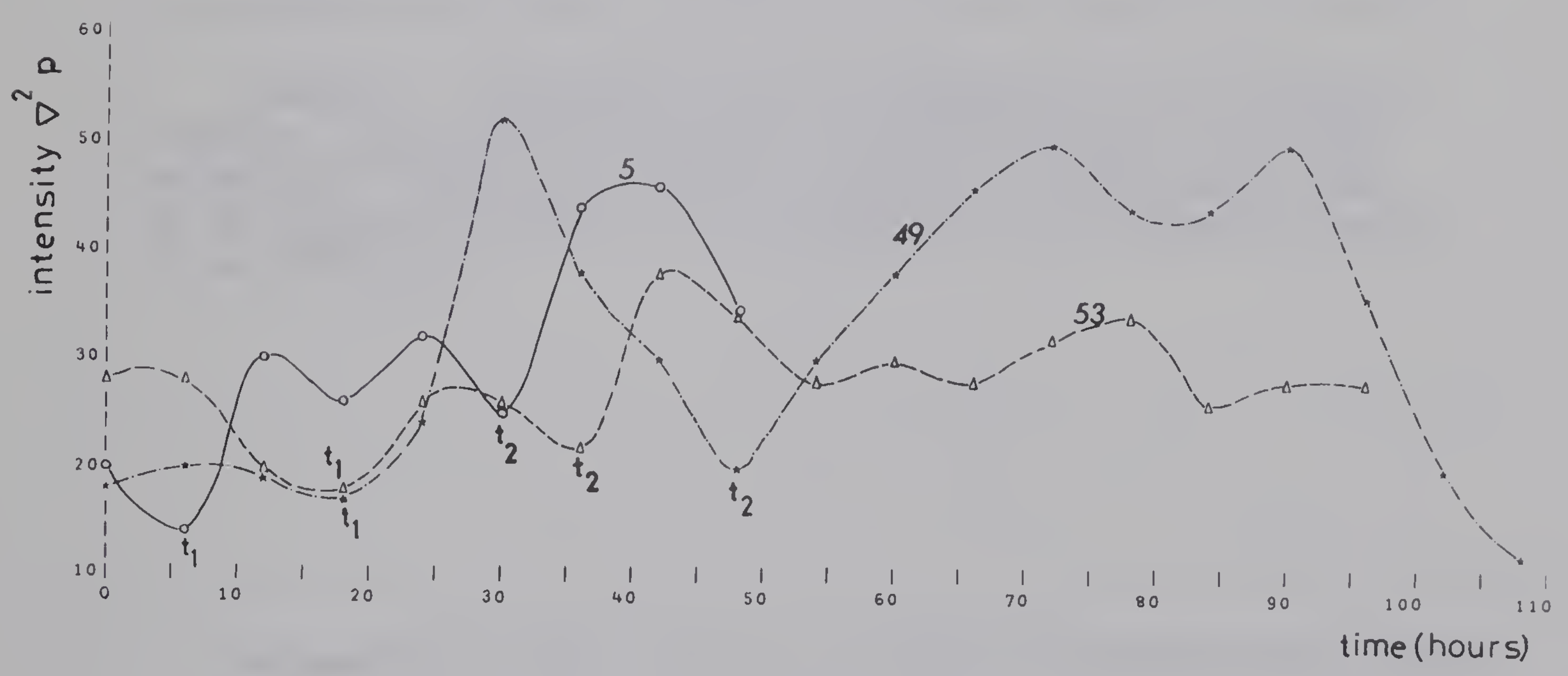


Fig. 10. Plots of intensity (∇^2 p) versus time t for six developing cyclones. ∇^2 p is given in units of mb/(334 km)^2, and the time in hours. The upper set of curves shows the history of three type A_1 developments with single onset times t_o. The lower curves show the development of three cyclones of Type A_2 with two periods of intensification beginning, respectively, at t_1 and t_2. The curves are labelled with the case numbers of the cyclones analyzed.

$\nabla^2 P \equiv J$ Intensities	Frequency			
	With trajectory		Local	Sum
	Single max. J	Double max. J	B	
58 - 38	A_1 13	A_2 10	0	23
37 - 30	13	11	4	28
29 - 26	22	7	2	31
24 - 22	C 17		D 9	26
21 or less	16		22	38
Total	109		37	146

Table 1. Frequency of lee developments in the Canadian Rockies, classified according to maximum intensity and type of motion. The block letters in the body of the table refer to the type of cyclones. The intensity ∇^2 p is given in units of mb/(334 km)^2.

Referring to Figs. 8 and 11, it may be seen that the maximum frequencies of the initial formation of cyclones occur in the lee of the principal mountain systems. The maximum in the lee of the Southwestern Alberta Range coincides with the location of the maximum determined by Petterssen (1950, 1956) shown in Fig. 2. However, the two other frequency maxima near Fort Nelson and Norman Wells do not appear on Petterssen's map. These discrepancies may be the result, in part, of the different size of the sampling areas. In the present study the frequencies of cyclogenesis were counted in a grid of square

sampling areas 1.5 by 1.5 degrees of latitude in size. Petterssen, on the other hand, used squares of 100,000 km^2, or approximately 3 by 3 degrees of latitude in size. Furthermore, Petterssen took the Historical Weather Map Series for the period 1899-1939 (one chart per day) as the basis for his analysis, whereas the four daily six-hourly synoptic maps were used in this study. In addition, many weak disturbances analyzed on the larger scale (1:12,500,000) maps can not be found on the Historical Maps (1:50,000,000). With respect to Fig. 11, it should be noted also that cyclogenesis occurring in Montana was included in the count, while all cyclogenesis initiated some 900 km to the east of the Continental Divide was not considered.

In Fig. 11, it will be noticed that frequencies in Southern Alberta, in the lee of the Southwestern Alberta Range, are almost double those occurring in the lee of the Northern B.C. Range and the Mackenzie Mountains. There are actually two maxima in Southern Alberta, but since they would probably merge if a larger data sample were analyzed, it may be preferable to consider them as a single maximum. But it seems that the maxima in the lee of the Mackenzie Mountains and Northern B.C. Range would retain their separate identities even if large data samples were used.

A high proportion of cyclone development is clearly and intimately associated with the three principal barriers, although a few cases of cyclogenesis do occur 500 or more kilometers to the east of the Divide. Fig. 12 shows the sites of first detection and distribution of all cases of cyclogenesis classified as Types A_1, A_2 and Group C; Fig. 13 is a similar plot of cyclogenesis of Type B and D. (The frequency isopleths in Fig. 11 are based on all 146 cases of lee cyclo-

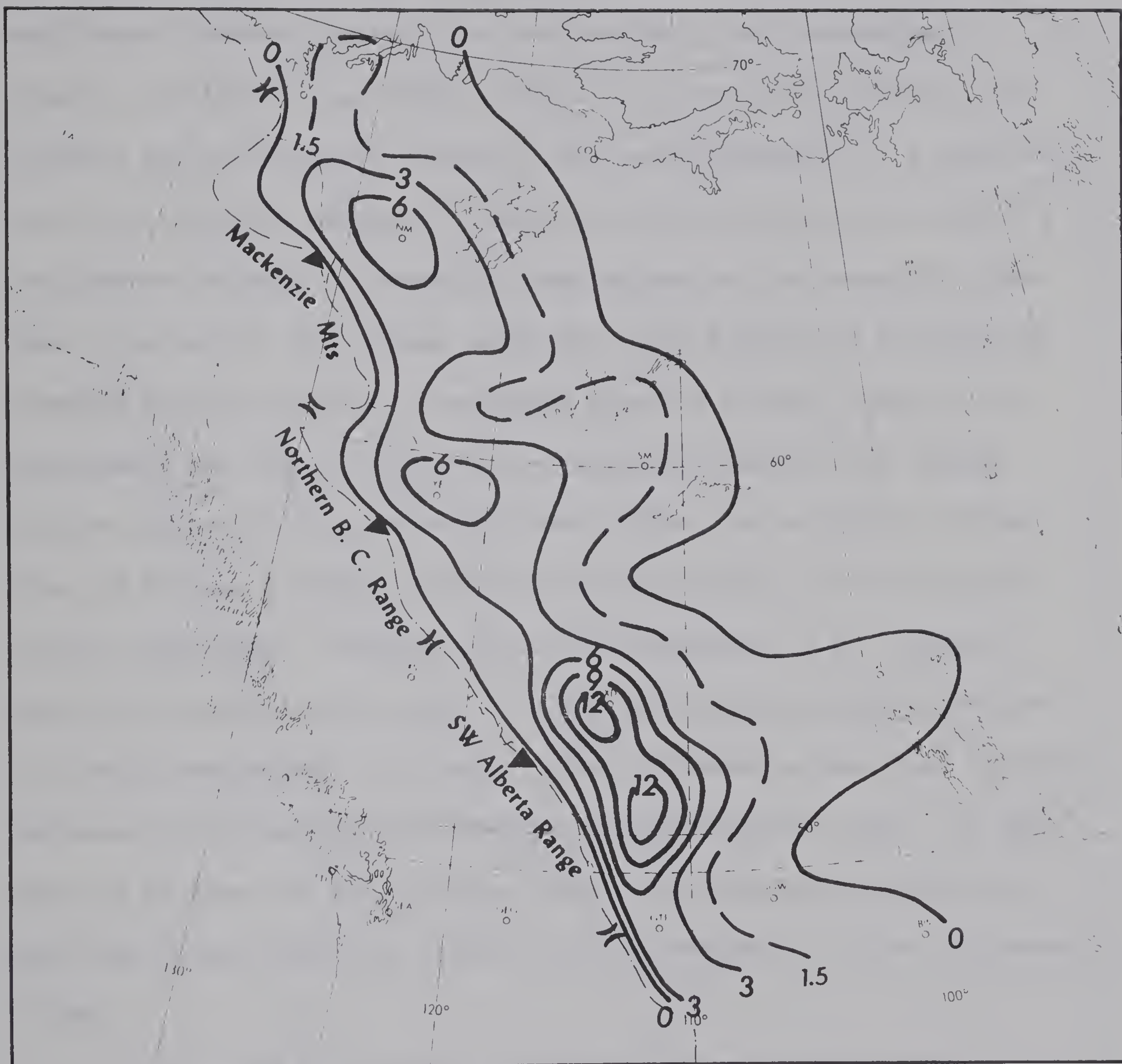


Fig. 11. Frequencies of cyclogenesis in the lee of the Canadian Rocky Mountains for 1958, counted in sampling areas 1.5 by 1.5 degrees of latitude in size.

genesis, the combined total plotted in Figs. 12 and 13.)

The distance between the centre of the frequency maximum and the Divide for the Alberta lee cyclones is of the order of 200-250 km. When denser network observations are available and intermediate-synoptic or hourly large-scale charts are used in the analysis, this distance may be decreased slightly, but not appreciably -- a problem which was carefully examined. However, the location of the site of cyclogenesis appears to depend to some degree on the strength of the upper circulation, for it was noted that lows frequently developed at somewhat greater distances downstream from the Divide, whenever the development was associated with a strong cross-barrier jet stream. Cyclones which first appeared far down stream, say at 500 to 700 km from the Divide, probably failed to develop earlier, either because certain conditions necessary for lee cyclogenesis, e.g., adequate vorticity advection were absent, or the generative processes operated at a much reduced rate. Of course, some of these cyclones may have been initiated by mechanisms not dependent on cross-barrier flow. In any case, at no time did such systems satisfy the criterion requiring at least one closed sea-level isobar in close proximity to the Continental Divide.

It should be noted that Mackenzie[1] lows occurred more frequently in the warm months. The causes might be sought in the northward shift of the jet stream and polar-frontal zones in the warm season. Supporting evidence for this is given in the Appendix of this

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[1]The cyclones developing in the lee of the principal ranges will be hereafter referred to as Mackenzie lows, (B.C.) North Range lows, and Alberta lows, respectively.

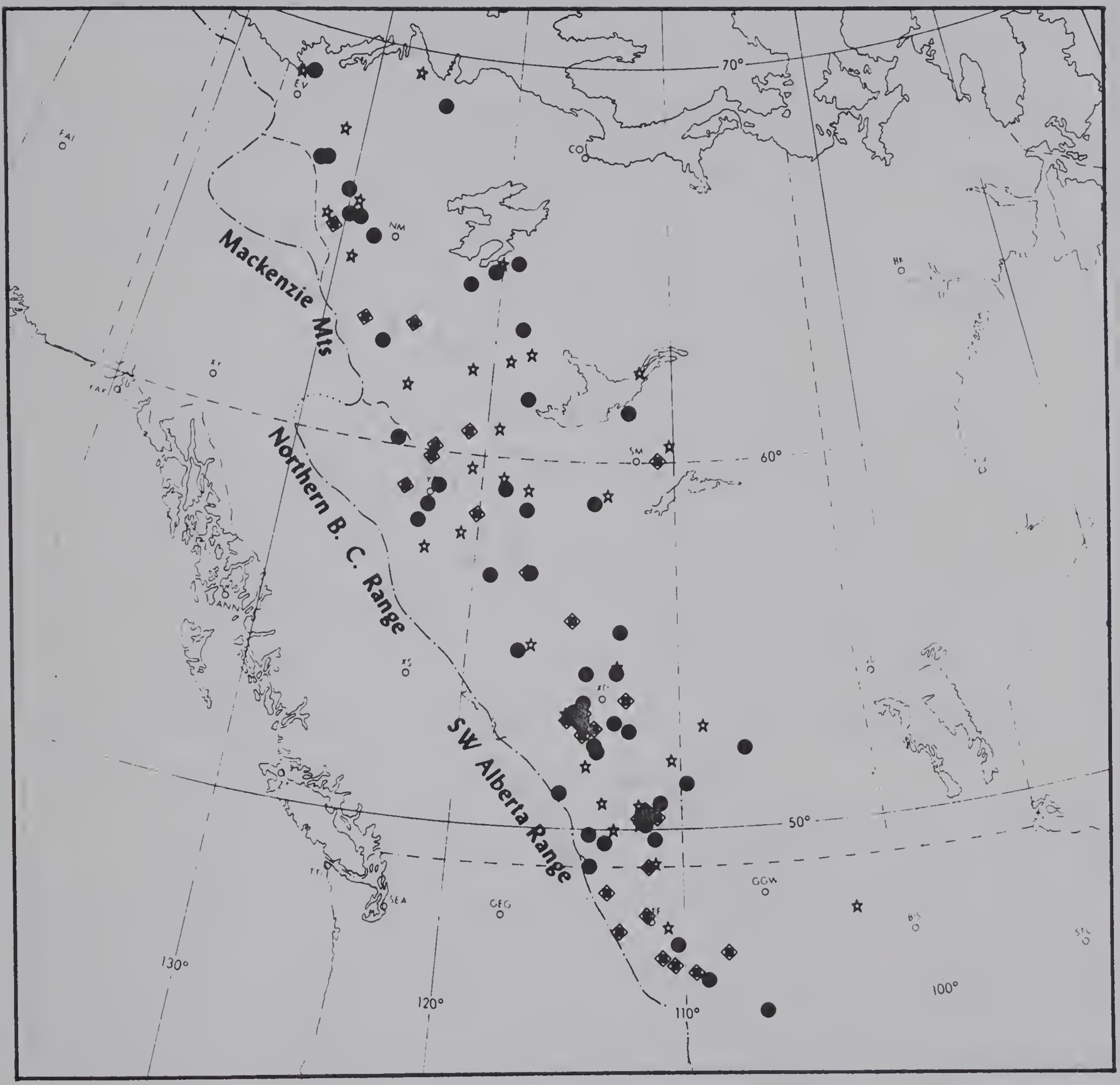

Fig. 12. Sites of cyclogenesis of migrating lows with long, well-defined trajectories. The sites of first detection are marked as follows: Type A_1 ● Type A_2 ◈ Type C ☆

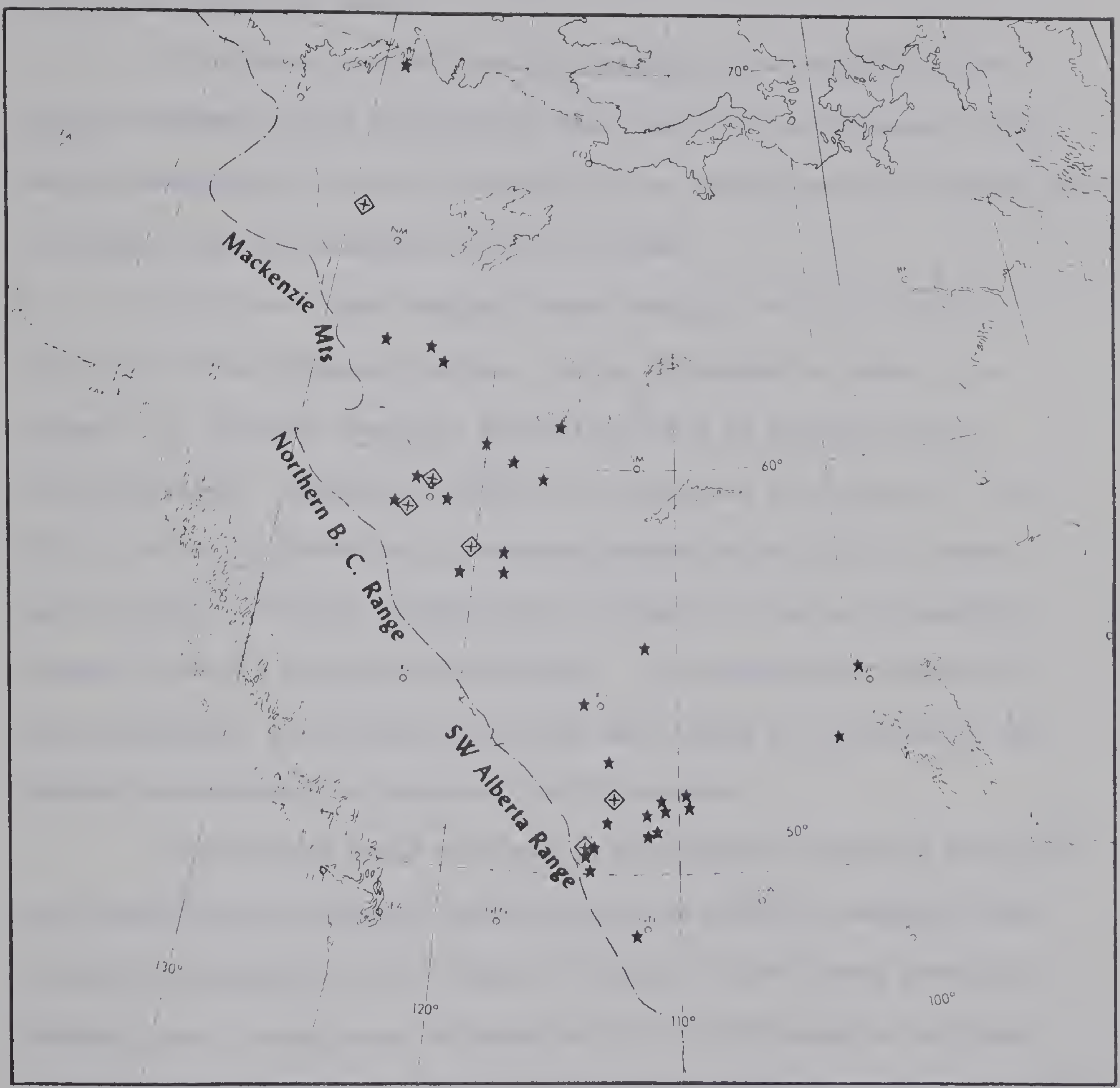

Fig. 13. Sites of cyclogenesis of quasi-stationary, local lows with short, poorly-defined trajectories. The sites of first detection are marked as follows: Type B ⊕ Type D ★

study. No attempt was made to construct maps of monthly frequencies, since such a classification would require data for at least three, and preferably five years.

It is also apparent that, in general, the Mackenzie lows acquire somewhat lower intensities than the Alberta cyclones. This seems reasonable, since the strength of jet stream over the arctic area is weaker, and the mountain barrier is lower.

It is not clear whether these results, in particular the location of the frequency maxima, can be explained in terms of a primary or, perhaps, dominant mechanism, such as synoptic-scale mountain waves. Radinović (1956b), who examined cyclogenesis in the lee of the Alps, found two pronounced maxima in the Gulf of Genoa and the Gulf of Venice, respectively, linked by a zone of somewhat lesser activity, through the Po Valley. By reanalyzing Radinović's block diagrams, the distance from the main ridge of the Alps to the maxima was estimated to be about 200 kilometers.

Considering these findings in the light of existing synoptic-scale mountain wave theory, some case can be made for mountain-wave induced cyclogenesis on the basis of Queney's (1948) wave profiles. However, more recent work by Queney et al. (1960) seems to mitigate against this mechanism in that it limits the maximum wave length of lee waves to about 70 kilometers, a mere third of the distance necessary for possible lee wave cyclogenesis south of the Alps and east of the Canadian Rockies. Not many references to work on synoptic-scale lee waves in the 200 kilometer range have been encountered in the course of this investigation, except for the theoretical studies of Dirks et al. (1967), Eliassen (1968), and Eliassen and Palm (1960) who consider

long, gravity-type mountain waves in the range of 100-1000 km.

In the absence of a general theory of synoptic-scale lee waves, the conclusions to be drawn from the empirical evidence presented so far must be tentative. *It is clear, however, that the great majority of the cyclonic systems which appear in the lee of mountain ranges are initiated by barrier-dependent processes. It is also evident that the frequency maxima of cyclogenesis in the lee of the Canadian Rockies are associated with the three principal ranges, and not with the low ground of the adjoining major passes.*

3.7 Tracks of Sea-Level Cyclones

Figs. 14 to 18, the five maps following page 66, are abstracts of the complex orographic history of the 146 cyclones classified and discussed in the two previous sections. Plotted on the maps are the tracks and sites of first and (where applicable) second intensification of all five cyclone classifications.

Attempts to determine the principal mean tracks of the cyclones were abandoned when it became clear that ten or more tracks would be necessary to describe the pattern with at least a marginal degree of fidelity. In any case, such charts of mean tracks seem to be of limited usefulness for prognostic purposes. However, tracks toward the southeast, east and northeast were found to be the most common, the directions most favoured, and, of course, well known. In a paper by Klein (1956) the principal cyclone tracks in northwestern North America are shown running toward the southeast. The tracks are given with but little explanatory comment, and no reference is made to other principal tracks. However, Queney (1948), and Hess and Wagner (1948)

have given clear and detailed explanations of the trajectories of air particles over mountains.

Almost all cyclones, shown in the following five maps, travel towards the north and fill as they approach the continent or, more properly, the high terrain of the Western Cordillera, despite the presence of a generally westerly flow aloft. This is further evidence that surface cyclones can not readily cross a broad mountain range, an observation known to and explained by Shaw (1906), referred to earlier in section 2.4.1. However, a small number of intense fast-moving cyclones associated with upper troughs were found to move bodily across the mountains without the occurrence of a distinct process of cyclolysis being evident on sea-level maps. In order to give a clear view of the initial positions and the sites of sudden intensification, the tracks are shown as discontinuous over the Continental Divide. Such tracks are composed of a relatively straight section near the Divide and approach close to the Divide in those figures.

In the lee of the Rocky Mountains, the great majority of tracks have a southerly component in the first few hours after cyclogenesis. It appears that the effect of slope on cyclonic movement operates most efficiently on the lee slope of a broad range of mountains. According to the Rossby equation (11), as the depth of an air column increases in the lee of a barrier, the trajectory of the air must acquire strong cyclonic curvature. Consequently, *lee cyclones will initially tend to move in a southerly direction along cyclonically curved trajectories while they are on the lee slope of a mountain range.* This should be compared with the earlier observation that cyclones track northward on the windward side of a range.

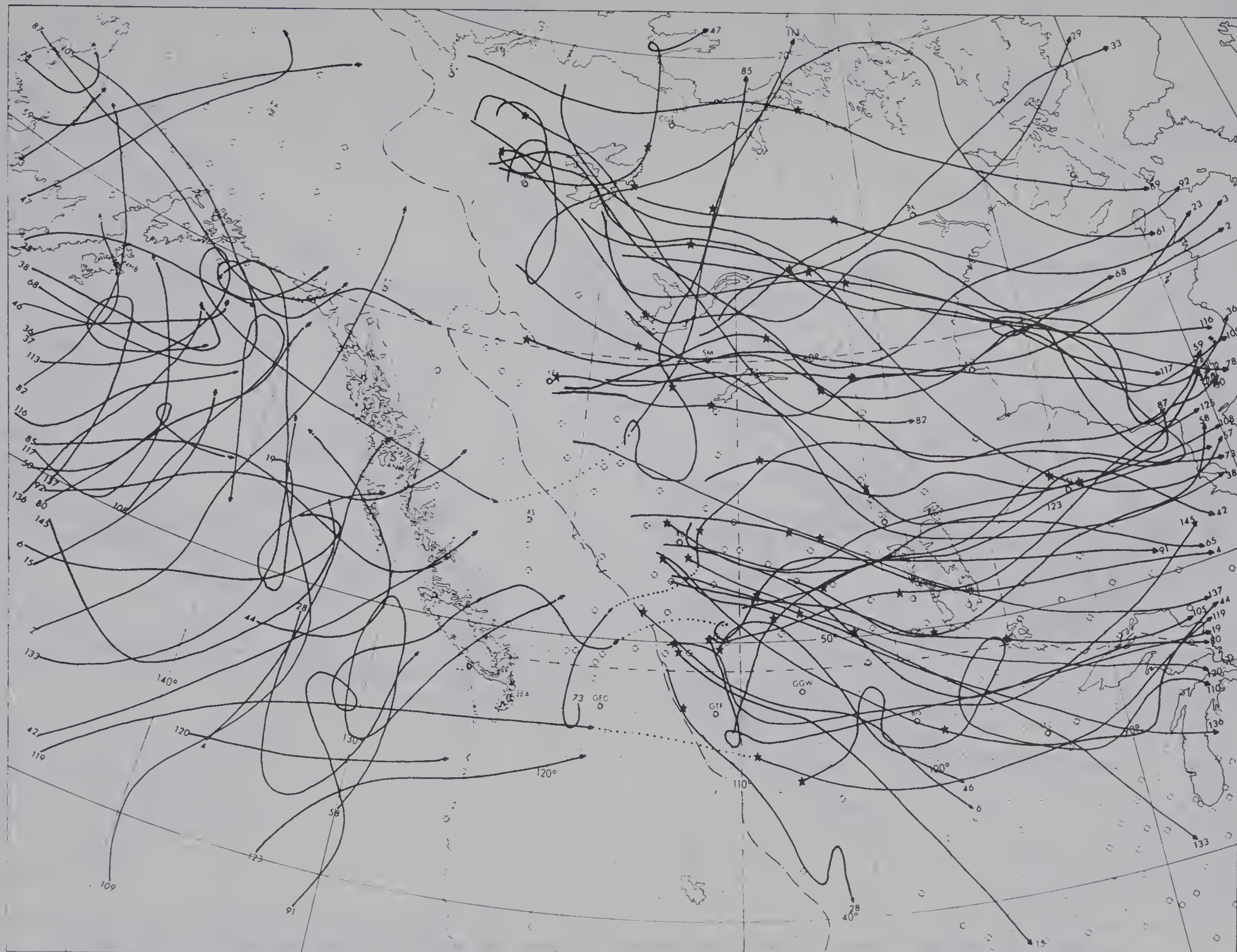

Fig. 14. Tracks and origins of lee cyclones of Type A_1 and tracks of their parent lows. The stars (★) denote the sites of sudden intensification at time t_o. The numbers identify the tracks of individual cyclones.

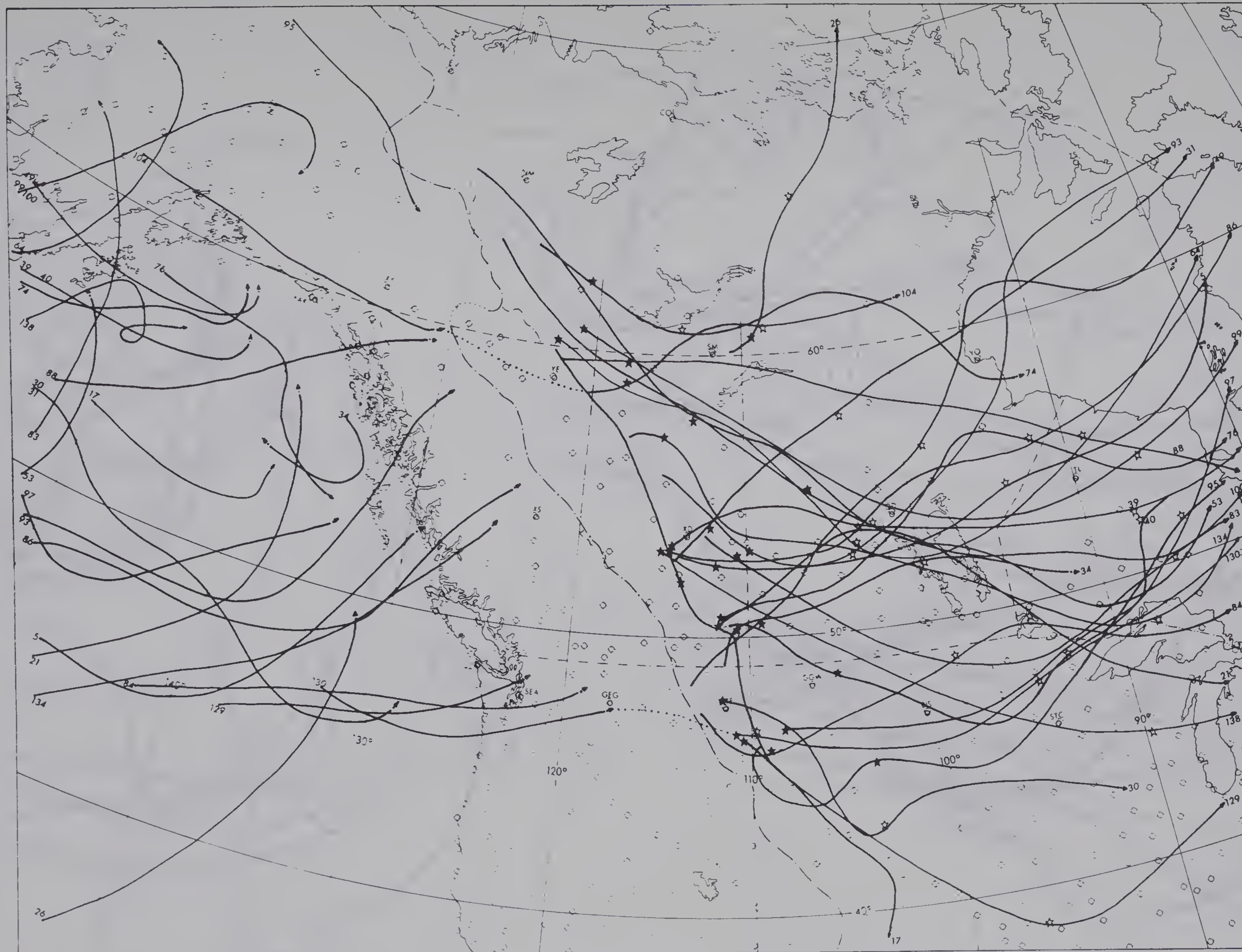

Fig. 15. Tracks and origins of lee cyclones of Type A_2 and tracks of their parent lows. The solid stars (★) denote the sites of first intensification and the open stars (☆) second intensification, at the corresponding onset times t_1 and t_2. The numbers identify the tracks of individual cyclones.

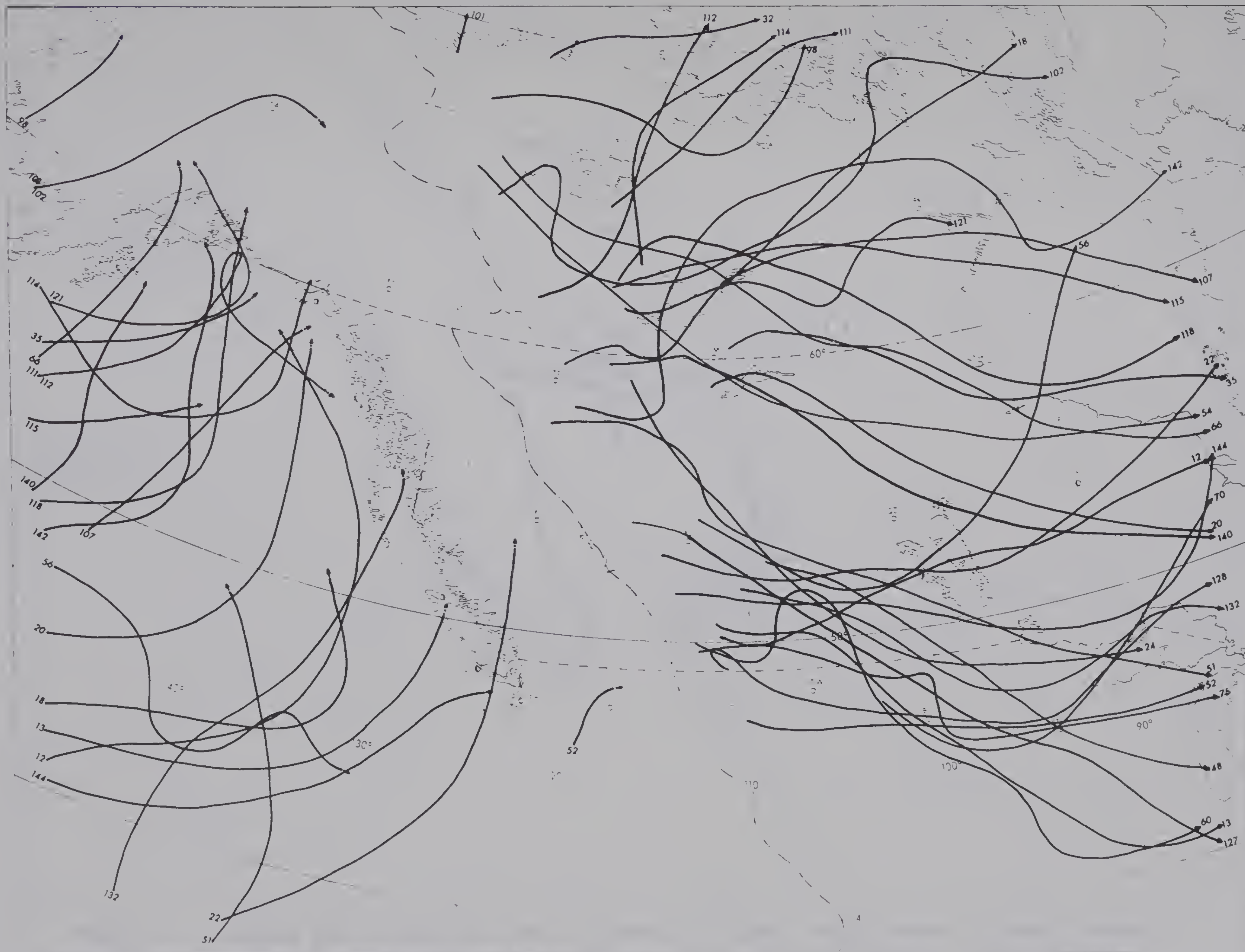

Fig. 16. Tracks and origins of lee cyclones of Type C and tracks of their parent lows. The numbers identify the tracks of individual cyclones.

Fig. 17. Tracks and origins of lee cyclones of Type B and tracks of their parent lows. The stars (★) denote the sites of intensification at time t_o. The numbers identify the tracks of individual cyclones.

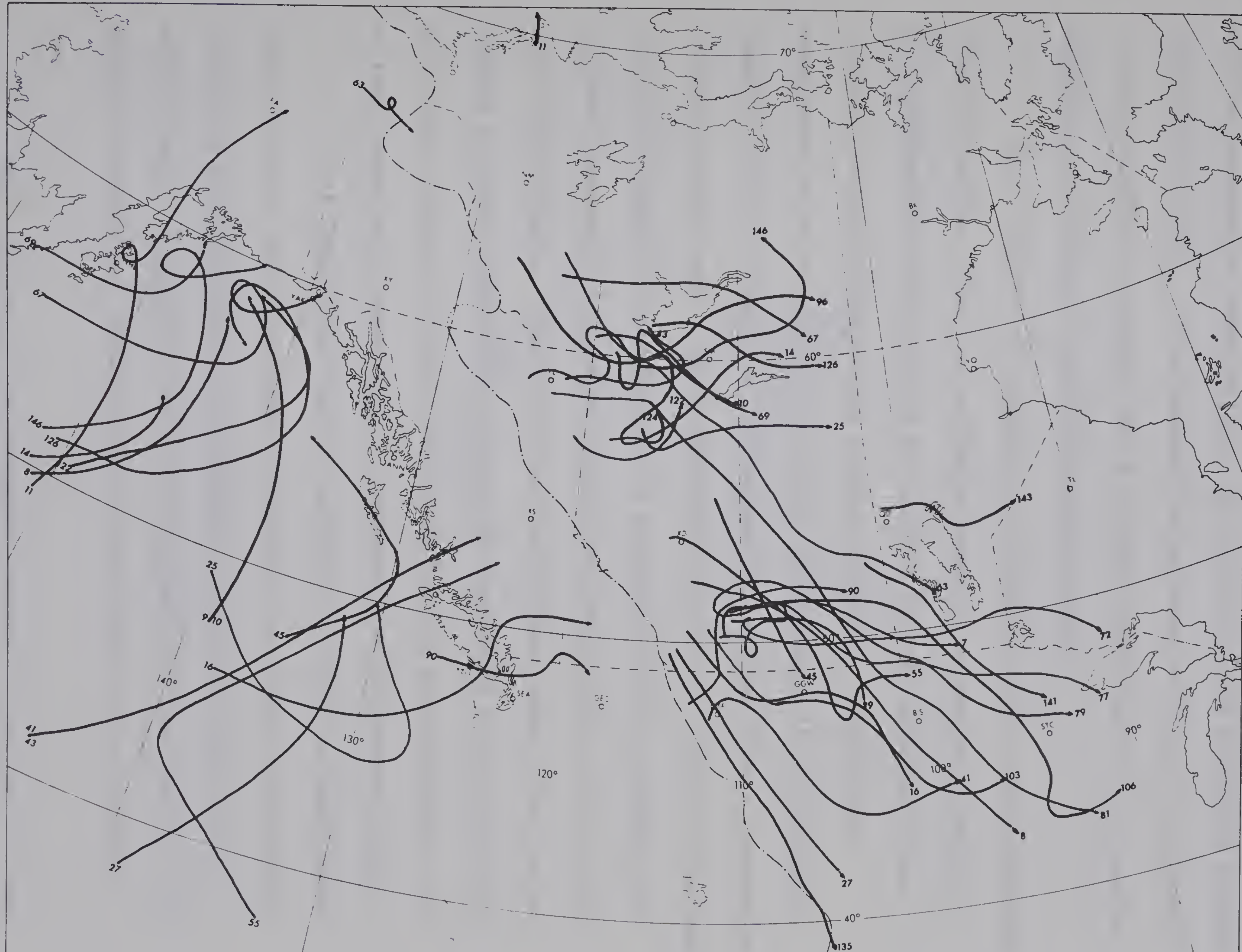

Fig. 18. Tracks and origins of lee cyclones of Type D and tracks of their parent lows. The numbers identify the tracks of individual cyclones.

As the slope decreases and the terrain flattens out, the depth of the air column will no longer increase, and the motion of the cyclone will be governed by other effects, in particular the occlusion process and latitude effect. This tendency is apparent in Figs. 14, 15 and 16, in the vicinity of longitude 80° W, another empirical fact in general agreement with Rossby's ideas on cyclone motion[1].

Fig. 19 shows an example of cyclogenesis on a quasi-stationary front on the lee slope, for a typical synoptic situation (case 93). It is noteworthy that about 49 per cent of all the cyclones formed initially on such pre-existing frontal surfaces. A cyclone was considered to be the result of frontal cyclogenesis if it developed roughly within a distance of two degrees of latitude from a previously existing frontal zone. The remaining 51 per cent of cyclones developed initially without the association of a distinct front, although lee cyclones frequently may become frontal cyclones as they move out of their source region. However, about half of the cases of lee cyclogenesis occurred essentially outside the frame-work of the classical polar-front theory, and other mechanisms have to be considered, such as local baroclinic zones, and the hypothesis of Petterssen (1955).

It is well known that the analysis of fronts in mountainous areas is not an easy task; some difficulties are always encountered because of irregularities in the thermal structure, various local effects, etc. In practice, due to the pressure of time and lack of adequate information, maps analyzed in the course of the routine

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[1]This subject will be discussed further in sections 3.9 and 5.1. See also Petterssen (1956) page 241.

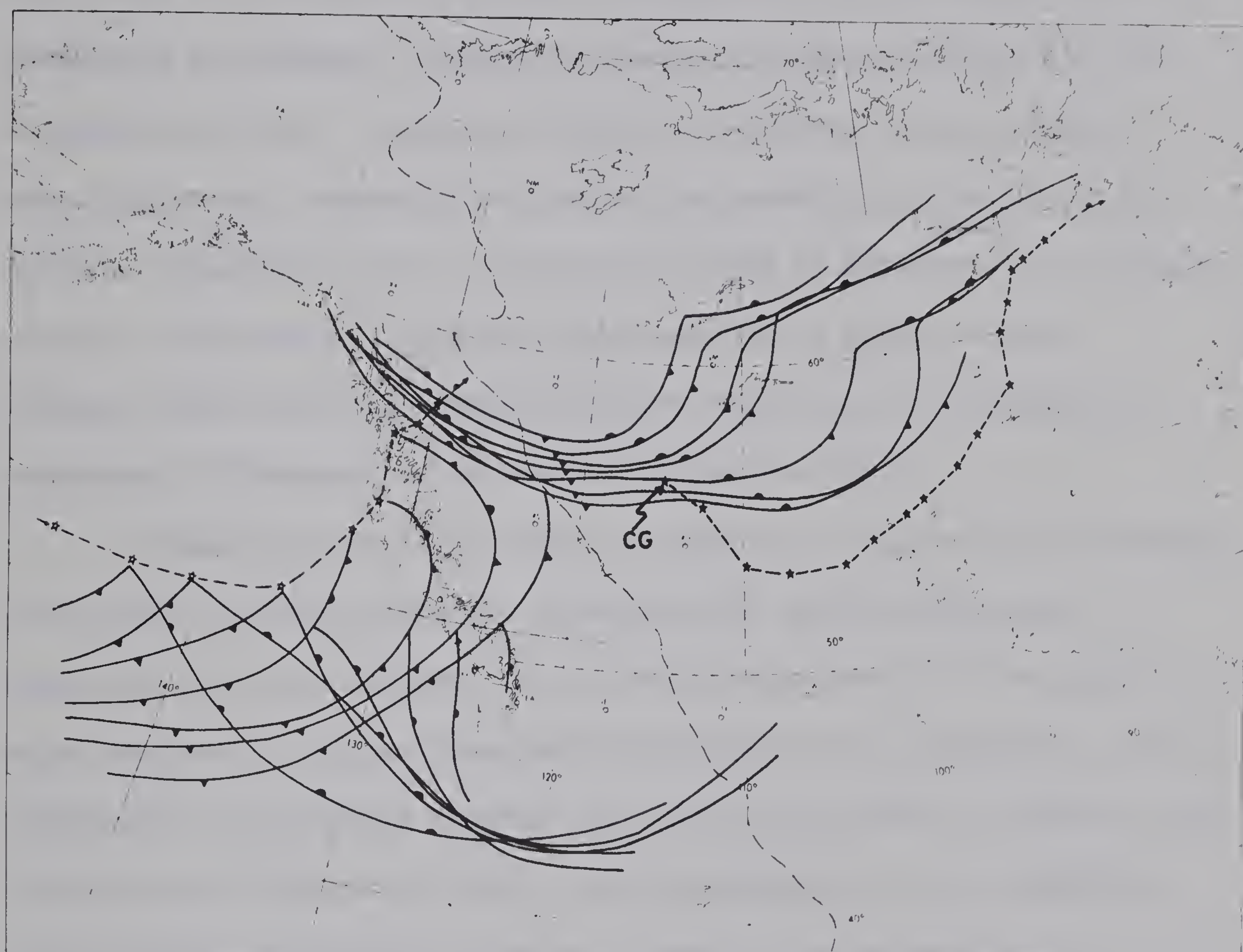

Fig. 19. A typical cyclogenesis on the arctic front (case 93), in the period of 19 to 23 September, 1958. For the sake of clarity, positions of the arctic front are not shown after cyclogenesis (CG) has occurred. Stars (★) mark the subsequent track and six-hourly positions of the cyclone.

operation of a weather office are often incomplete, if not incorrect. Such maps, when used later in synoptic studies, may have to be re-analyzed, but some errors will remain, and a "definitive analysis" will always be an elusive goal. Some of the maps used in this study, when found to be wanting in detail, consistency or continuity, were reanalyzed and amended, insofar as that was possible without the use of additional data. Though not likely to be shown on maps, highly baroclinic zones frequently do develop and persist in the lee of the Rockies, especially when a strong anticyclone is dominant over Central Canada. Furthermore, the Rocky Mountains form a great, natural barrier, which not only hinders the free flow of air, but serves to accentuate differences in air masses across the Divide.

Regarding the 51 per cent of cyclones not directly associated with fronts, it is, of course, possible that certain baroclinic characteristics are inherent in the air flowing over the lee slope, which may tend to create favourable conditions for cyclogenesis. This possibility was explored briefly in section 2.5; there is little in the literature to support this idea, and no reference to it in mountain wave theories and synoptic studies. However, the atmosphere is frequently said to be more or less in a "state of baroclinicity", since it is easily disturbed by several kinds of influences, such as orography, shearing motions, various diabatic effects, etc. As mentioned earlier in section 2.4.1, V. Bjerknes et al. (1911) considered such air mass disturbance ensuing from flow over mountains. Thus even if the isothermal and isobaric surfaces coincide in a region of the atmosphere, this barotropic state, called "incidental barotropy" by Petterssen[1], will be rather short-lived, as a result of frictional

[1] See Petterssen (1956) page 107.

and thermal effects, etc. Furthermore, the incoming "disturbed" winds from the windward side of a mountain range will effectively prevent the maintenance of the barotropic state. Therefore, the atmosphere may be assumed to be baroclinic whenever there is appreciable downward motion in the lee of a range of mountains. Large shearing motions alone will probably not produce a distinct discontinuity on the steep slopes of mountains and, in any event, it would be difficult to locate such discontinuities in practice, from the limited information provided by a synoptic network.

There is a "quasi-frontal" effect of mountains which may be of some importance in lee cyclogenesis under certain conditions, discussed by Hage (1957) and by Newton (1958: see Reiter; 1963). As noted earlier in section 3.2, the smoothed or average slope of mountains is of the same order of magnitude as a typical atmospheric frontal slope. Prolonged upslope motion on this fixed "frontal" surface will also disturb the isobaric-isosteric layering of an air mass, creating solenoids and baroclinic conditions. It seems reasonable then to conclude that the 51 per cent of cyclones which were not associated with distinct synoptic-scale fronts, developed in localized hyper-baroclinic zones, in accordance with one of the postulates in Petterssen's hypothesis. These possible mechanisms will be examined in more detail in chapters 4 and 5.

3.8 The Initial Formation of Lee Cyclones and the Diffluent Upper Pattern Quasi-Normal to the Canadian Rockies

An attempt was made to apply some of the ideas of synoptic-scale barrier wave theory to lee disturbances, and to examine the implications of "classical" theory to the sample of 146 lee cyclones. As will be recalled from the foregoing chapter, all barrier wave theories basically assume an air flow of considerable speed and depth normal to the ridge of mountains. These are rather stringent requirements not often satisfied under the less than ideal, but real conditions of day-to-day synoptic flow and development. For synoptic purposes, Scherhag's (1934, 1937) ideas (referred to earlier in section 1.4) were found to be the easiest to apply in practice.

In the present study, these diverse ideas, as well as some basic fluid-dynamical principles, are tentatively used to examine the sample of lee cyclones. But it should be emphasized that the tests apply mainly to *the initial formation (rather than intensification) of lee cyclones* associated with the three principal mountain systems of the Canadian Rockies. On the basis of mutually consistent theoretical and empirical considerations, the following criteria were established as describing conditions either favourable to, or observed at the time of lee cyclogenesis:

(i) A diffluent 500-mb contour pattern was present over any one of the three principal mountain systems.

(ii) The flow pattern was normal to, or nearly normal to the ridge of any one or all of the three principal ranges.

(iii) A jet stream or a core of relatively strong winds[1] was crossing the mountains.

(iv) The distance from the Continental Divide to the nearest upstream trough was large, i.e., 500 km or more.

In the course of the present analysis, all lee cyclogenesis on sea-level charts was examined for possible association with the 500-mb flow patterns for at least 24 hours prior to the initial development, to 24 hours thereafter. Some of the original working charts needed to be reanalyzed in order to remove various inconsistencies from the synoptic sequence of flow patterns.

The angles at which diffluent 500-mb contours cross the Continental Divide were roughly estimated on each 12 hourly 500-mb chart, in order to select all cross-barrier flows satisfying criterion (ii). If the angle was within the range of about 70-90 degrees, the 500-mb flow was considered to be a "perpendicular flow"; a more detailed classification is hardly warranted in view of the sparse upper-air sounding network. On the basis of this analysis, the 500-mb flows associated with the surface cyclones classify as follows:

Angle between the Continental Divide and the 500-mb contours	Percentage of flow pattern aloft associated with the 146 cyclones
90 - 70 degrees	ca. 80%
65 - 50 degrees	ca. 7%
45 - 0 degrees	ca. 13%

[1]Those winds were not strong enough to satisfy the definition of the jet stream, given in the Glossary of Meteorology of the American Meteorological Society.

Furthermore, all cross-mountain jet streams and regions of relatively strong 500-mb winds over the study area were examined with regard to criterion (iii). It was found that about 79 per cent of the developments were initiated under jet streams or concentrations of relatively high 500-mb winds. This result is in good agreement with the studies of V. Bjerknes et al. (1911), Riehl et al. (1952) and Reiter (1963), referred to before in section 2.1. *It is apparent that most of the lee cyclones in the Canadian Rockies occur with strong upper air flows which cross the ridges of the three principal mountain systems at angles near 90 degrees. This in turn indicates that the vorticity is largely generated at near normal incidence by the incoming strong winds to the lee*[1]. As far as could be ascertained, these conditions seemed to be completely absent in the remaining one-fifth of the cases.

Much emphasis was placed on the examination of flow patterns under the constraint of criterion (i). *It was found that about 83 per cent of all cases of lee cyclogenesis in the Canadian Rockies occurred under diffluent upper flow patterns, preferably several hours after an upper ridge had passed the Continental Divide.* On the other hand, only about 12 per cent of the cases were found under a parallel flow, while the remaining 5 per cent were associated with confluent patterns. Moreover, disturbances initiated under these "deviant" patterns of flow were, on the average, much weaker. It seems plausible that these latter percentages could be somewhat decreased with the addition of more observations in the presently data-sparse mountainous regions. However,

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[1]Vorticity production of this kind is readily demonstrated in wind tunnel experiments.

a perpendicular diffluent pattern was found on the 700-mb level for some of these exceptional cases.

When examining cyclogenesis with regard to criterion (iv), it was found that lee cyclone initiation did occur independently of, i.e., without the definite, close association of an upper trough or positive vorticity advection upstream. A number of vorticity charts were drawn up in order to study the effect of advection. It will be recalled that the absolute vorticity is small in an anticyclonic ridge aloft, and that the vorticity decreases rapidly with the approach of an upper ridge. The position of the maximum of positive vorticity advection at the 500-mb level was found to be, in general, between a ridge and a trough somewhere in the area of strong gradients of the 500-mb contours, preferably associated with a jet stream. On the whole, separation distances[1] for the majority of cyclones examined ranged upward from about 500 to a few thousand kilometers, in the earliest stages of lee cyclone development.

The great majority of cyclones were initiated in situations meeting the requirements of all four criteria. This includes almost all local cyclones of Types B and D, and most trajectory Type C cyclones. It should also be noted that initiation of lee cyclones of Type A occurred most frequently under the eastern edge of a diffluence pattern, and that intensification began commonly with the arrival of an upper trough to the east of the Continental Divide.

There are, of course, other orographic processes which may contribute to lee cyclogenesis. Thus, if the winds blow against a range

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[1]This term denotes the distance separating a surface low from upstream upper-air support, such as a trough or a cold low.

of mountains, a lee trough often develops as a result of vorticity production, in the manner required by the Rossby equation (11). Furthermore, as shown by Godson (1948) and Palmén and Newton (1969), pressure falls can occur without horizontal divergence over sloping terrain with downslope motion, because air is in effect removed vertically through the base of the subsiding column. These processes will be considered further in Chapters 4 and 5.

It should be emphasized that the time and the place of the first appearance of a cyclone in the lee were determined primarily by the diffluent pattern aloft. If southwesterly diffluent flow conditions were present over the Continental Divide, a lee cyclone could usually be found on the surface maps within the next 6 to 18 hours, and most often within 6 to 12 hours, but much depended on the relative movement of the diffluent pattern aloft[1].

Summarizing the foregoing, it appears that *a lee cyclone forms initially under a strong, diffluent upper-air flow perpendicular to a range of mountains, and intensifies with the approach of an upper cold trough.* The most favourable conditions for the initial formation of a lee cyclone (to the stage marked by the first closed isobar) were clearly those which satisfied all four criteria simultaneously.

Observations of the 700 and 500-mb contour patterns clearly show that initially parallel onshore flows tend to "fan out" laterally, and become diffluent flows, in the course of prolonged upslope motion over the Cordillera, on the way from the Pacific to the Continental Divide. But as shown by Hess and Wagner (1948), Petterssen (1956)

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[1]These aspects of cyclogenesis will be considered more fully in chapter 5.

and others, upslope motion over mountains produces vertical shrinking in the air columns, and downslope motion vertical stretching. These processes in turn are associated with horizontal divergence and convergence respectively. Thus it appears that the diffluent cross-barrier flow is the result of vertical shrinking and horizontal divergence. As noted earlier in chapter 2, the orographically induced process of stretching and shrinking of layers is implicit in the ideas of Rossby (1940), and proceeds in accordance with the ω - equation of Sutcliffe and Godart (1942)[1].

3.9 Tracks of Upper Cold Troughs and Intensification of Sea-Level Cyclones

For the consideration of upper air motions, the tracks of all 76 upper cold lows and troughs associated with lee cyclones are shown in Figs. 20 and 21. It will be seen that the general features of Figs. 20 and 21 bear close resemblance to the respective surface tracks of each type, although there are some differences in the tracks of the (Type A_1 and A_2) corresponding sea-level cyclones. Each position of an upper trough was identified and correlated with the position of the associated sea-level lee cyclone.

When analyzing the structure of upper lows, it was found that most of them tend to dissipate as they cross the Cordillera and the Continental Divide. It is of interest to note that *only about 26 per cent of the upper waves crossed the Divide with a closed contour on 500-mb charts; all others passed the Divide as cold troughs. This*

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[1]See Chapter 5 for details.

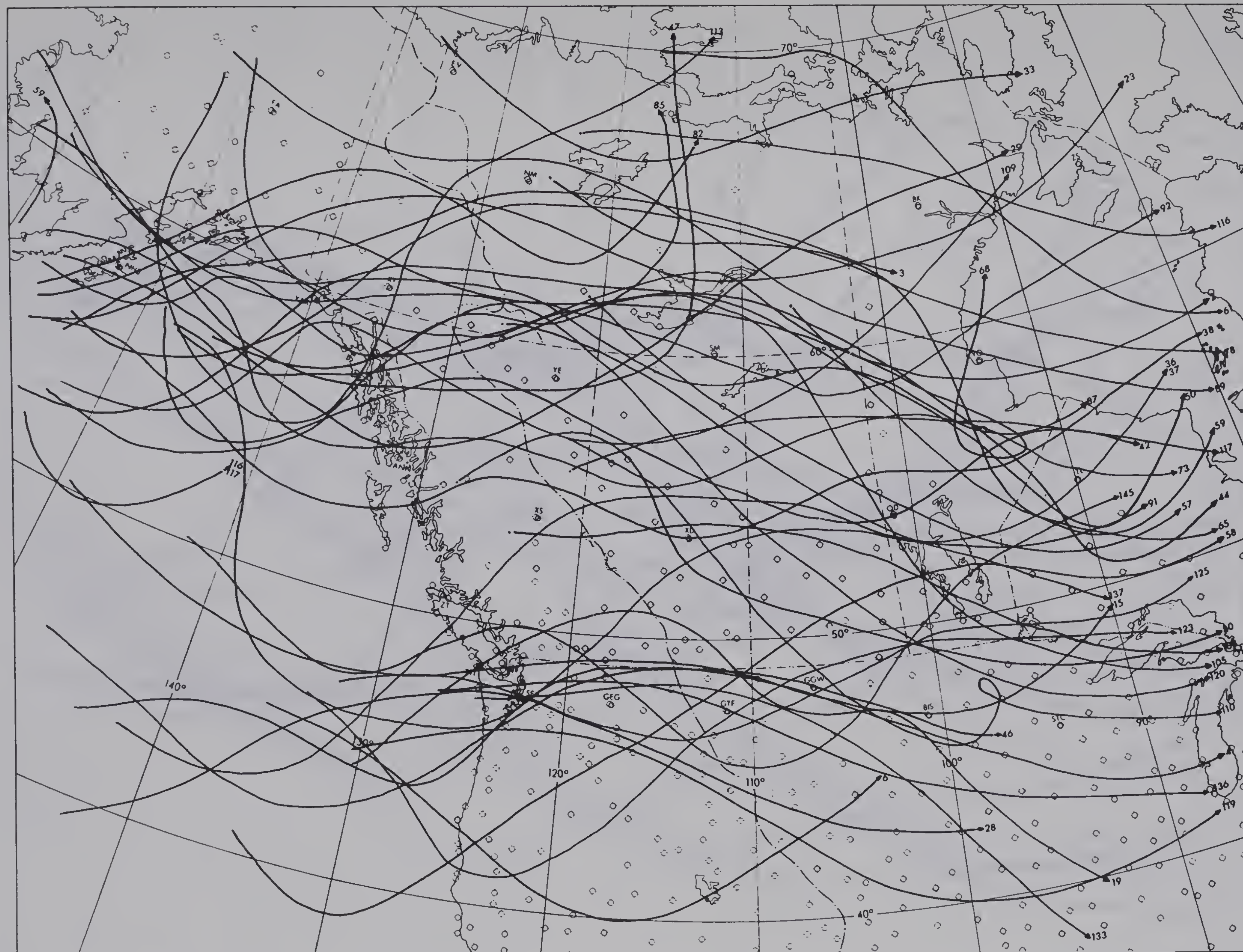

Fig. 20. Tracks of upper cold lows and troughs associated with lee cyclones of Type A_1. The numbers identify the tracks of individual cyclones.

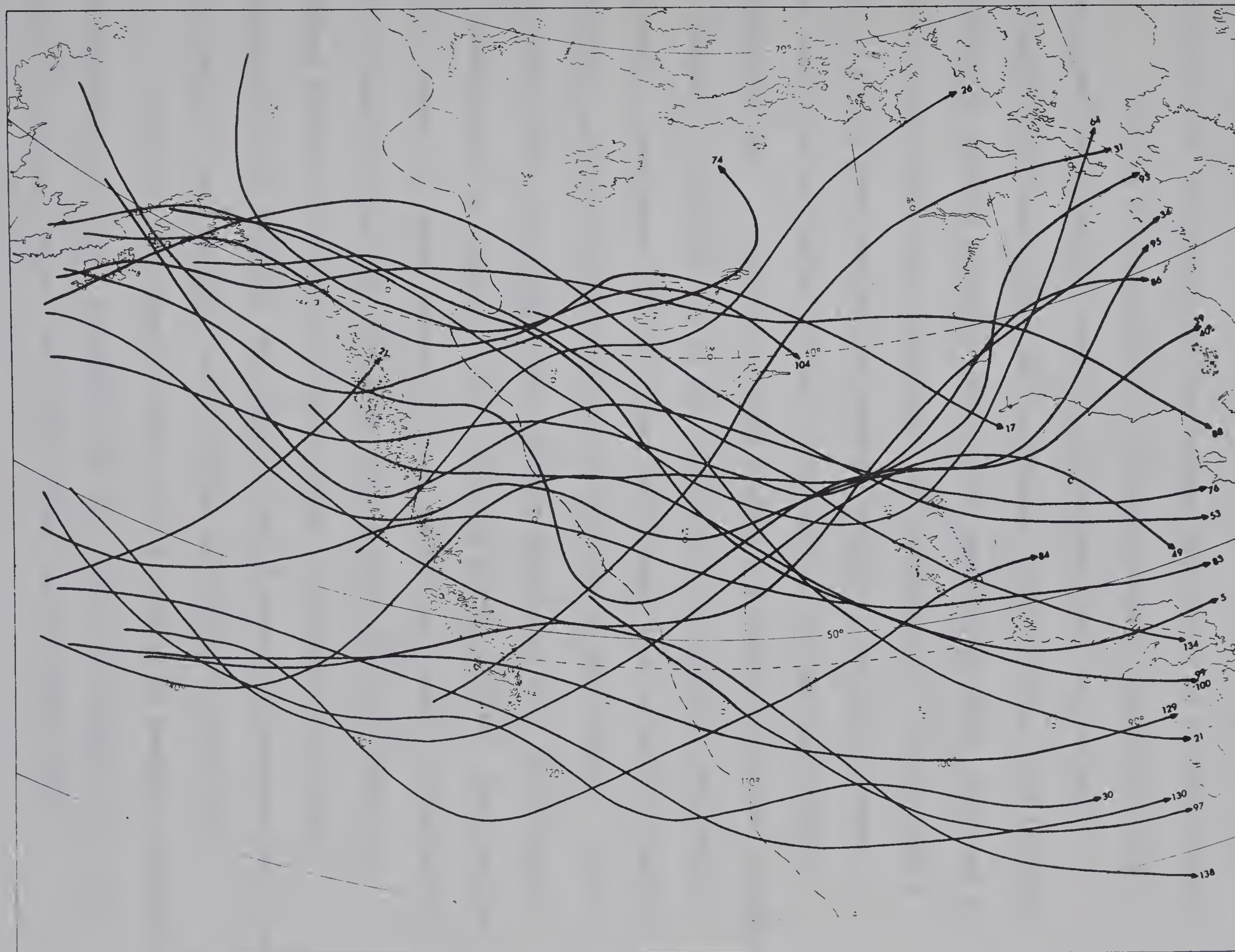

Fig. 21. Tracks of upper cold lows and troughs associated with lee cyclones of Type A_2. The numbers identify the tracks of individual cyclones.

obviously indicates that even 500-mb lows tend to be modified markedly by massive mountain barriers. These observations are in agreement with, and strongly supported by a study done by Hage (1957) who found that only 22 out of 61 upper cold lows passed the Divide as closed centres.

It will be recalled that the relationship between sea-level development and positive vorticity advection aloft was explored thoroughly by investigators of the Chicago school in the late fifties. Lacking the routine computer analyses of more recent days, they had to produce their own charts, piece-meal and laboriously[1]. When working with the 1958 data used in the present analysis, vorticity charts were again not available and had to be made up in the time-tested, "classical" way. A number of these charts are shown in the case studies of chapter 4. However, it is well known that a vorticity maximum is intimately associated with an upper cold trough and, moreover, positive vorticity advection usually occurs to the east of the trough. It is also known that such systems are closely associated with jet streams, as demonstrated by Riehl et al. (1952) and Reiter (1963). (See section 2.1.) Hence, it seems reasonable to identify the point of intersection of a trough line and the core of a wind maximum with the centre of maximum vorticity of an upper wave system. Although open to some objections, this approach was taken to locate vorticity centres and also the tracks of cold troughs on the 500-mb maps used in this study.

As it will be seen in Fig. 22, the centres of maximum vorticity and maximum wind do not coincide with the centre of an upper cold low. On occasion these centres are widely separated from each other,

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[1]Dr. Hage comments that "many a long hour was spent calculating vorticity and vorticity advection". (Personal communication)

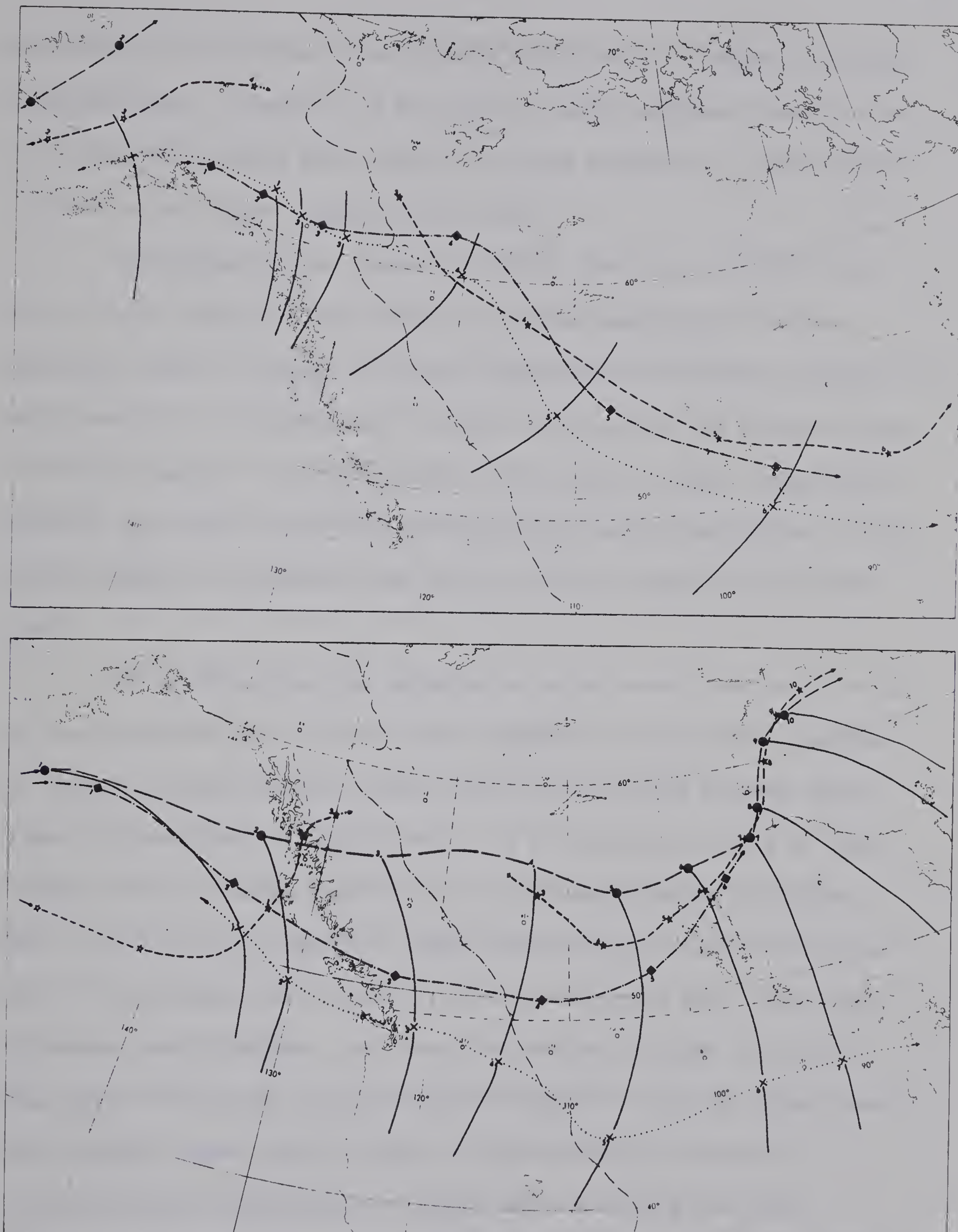

Fig. 22. Tracks of a typical upper cold trough (upper map, case 100) and an upper cold low (lower map, case 93), and their associated motion systems. The successive positions of the systems along their respective tracks are marked as follows:

● centres of cold lows   ◈ centres of vorticity maxima
★ centres of lee cyclones   ✪ centres of cyclones west of the Divide
× point of intersection of a trough with the axis of an isotach maximum

particularly in the case of an intense cold low or a trough of a very large amplitude. However, in the case of small amplitude waves it is found that the centres move usually in close proximity of each other, as shown in the upper diagram of Fig. 22[1].

Investigating the motion of 500-mb lows on consecutive 24-hourly charts, Hage (1957) deduced, with some smoothing, the four principal tracks of summer cold lows frequenting the western regions of North America. In attempting to extend this idea in the present study to 500-mb troughs, it was found that the tracks of upper troughs were somewhat more erratic and less concentrated, and the amplitude of the tracks seemed to be smaller than Hage's in the vicinity of the West Coast.

It is known that the majority of upper waves tend to curve to the north as they approach the Coast Mountains (as sea-level cyclones do) except for some erratic lows, and for fast moving systems which travel in relatively straight lines. In a few cases, tracks of upper troughs did not clearly dissipate on the windward before reforming again in the lee. In addition, Hage also noted that the lows did not behave in accordance with Rossby's theory. He found that orographic influences were prominent, and that the motions of upper cold lows were usually slow over the Northeastern Pacific, while an abrupt eastward movement began near the point of recurvature of the path. Acceleration and deceleration of upper waves crossing the Rocky

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[1] However, it should be noted that the great majority of lee cyclones initially occur usually well downstream in the northeastern or eastern part of such upper systems.

Mountains appear to involve another mechanism, to be discussed in chapter 5.

It was found that, on the whole, no cases of sea-level cyclones of Types B, C and D were associated with distinct upper cold troughs. However, referring back to sections 3.4, 3.5 and 3.6, it should be noted that the arbitrarily chosen limits of cyclone intensity do not determine exactly whether a surface low is associated with an upper trough. In the grouping of the present sample of lee cyclones, the difference in intensity between strong and weak cyclones is 2 mb per $(334\ \mathrm{km})^2$, a range which was found to be well-suited for the purposes of this study. It should be also noted that, in the study of Schallert (1962), none of the disturbances of Types B, C and D were associated with an advancing upper cold trough.

A sudden intensification is marked with the symbol of a star (i.e., the time at which the intensity begins to increase sharply) on each of the tracks of sea-level cyclones of Type A_1, A_2 and B in the previous Figs. 14, 15 and 17. The tracks and consecutive positions of each surface cyclone plotted on these maps had to be identified and associated with the tracks of the upper troughs in Figs. 20 and 21. From Fig. 20 (development Type A_1) it will be noticed that the positions of sudden intensification are widely scattered over the relatively flat terrain. The majority of cyclones intensified suddenly on the gently sloping surface some 300 to 500 kilometers from the Divide. But some of the fast moving cyclones, crossing the mountains through the generally lower terrain in the Alberta-Montana border region, began to intensify after upper troughs crossed the Divide, or while approaching it. During this phase of development, the separation distances between the upper troughs

and the surface cyclones at the beginning of this rapid intensification were still generally over 500 kilometers. Such cyclones attained their maximum intensity within 6 to 12 hours, if an upper trough crossed the Divide. This process of sudden initiation and rapid development may be interpreted in terms of the hypothesis of Petterssen.

It will be seen from Fig. 21 that the first sudden intensification (solid star) of Type A_2 cyclones occurred, on many occasions, closer to the Divide than the intensifications of Type A_1. This is in agreement with the findings of Schallert (1962) who noted that the mean position of rapid development of 6 cases of Type A_2 is relatively close to the Divide. No attempt was made in this study to define the mean position of Type A_2 cyclone development, since the individual positions of these systems varied greatly from case to case.

The first intensification commenced usually under a pronounced diffluent flow aloft, with the approach of a cold trough, especially when the trough was located west of the Divide. This first phase of development appears to depend markedly on orographic vorticity production. By contrast, the second rapid intensification (open stars) of Type A_2 cyclones occurred generally under an upper cold trough or with troughs a short distance upstream, i.e., with a small separation distance. Schallert called such an intensification a "normal development", mainly brough about by positive vorticity advection aloft. With continuing vorticity advection, the second period of sudden intensification leads usually to major cyclonic development. However, it should be emphasized that *almost* all lee cyclones, including the intense developments of Group A, were initiated within the requisites of the four criteria of lee cyclogenesis.

Referring back to Fig. 16, it may be also noted that most cyclones of Type C developed initially within the prescribed criteria, but then tended to travel great distances coupled with the upper flow, frequently under an isotach maximum. Considered as a group, such pressure systems were small in size, showed little growth, and little change in intensity.

As shown earlier in Figs. 17 and 18, the tracks and motions of Type B and D cyclones were confined almost entirely to their source region. Moreover, almost all these cyclones were generated beneath diffluent upper flows[1] in regions very close to the Divide. (See Fig. 13.) Most of these cyclones depend for their existence on structural changes of the diffluent patterns and, if upper air support in the form of troughs, jets, or vorticity advection is not available, they simply decay. Though small, these local disturbances are of considerable synoptic interest and, if of lesser importance, still integral parts of the general circulation.

While analyzing the surface charts and comparing the characteristics of the first and the second intensification of Type A_2 cyclones, an interesting fact emerged. The majority of cyclones underwent their second intensification over only gently sloping terrain, as may be seen in Fig. 15. These intensifications appear to be caused by non-orographic mechanisms, such as vorticity advection, diabatic effects, vorticity export from the Great Lakes, etc. Such processes are included in the development equation of Petterssen (1955). Furthermore, in many cases, the second intensification may be also identified with the second

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[1]The process is coupled with other effects, to be described more fully in chapters 4 and 5.

vertical stretching of layers, which occurs approximately 950 km from the Continental Divide, within the scale of the Queney (1948) model shown in Fig. 5. Of course, this theoretical model does not necessarily apply to the conditions of the actual atmosphere, and only qualified support can be given to this mechanism at the present time[1]. The question also arises whether the diurnal pressure wave contributes in any way to the intensification of migratory lee cyclones. In view of the significant pressure fluctuations reported by Haurwitz (1965) and Longley (1969), it seems plausible that the diurnal wave has some influence on lee development. In a study of long mountain waves, Eliassen and Palm (1960) also think that atmospheric tides (diurnal temperature variation, etc.) could contribute as "secondary energy-sources" to the generation of these waves, while orography acts as a "primary" source. As far as can be ascertained, this possible effect has not received much attention in the literature, but the entire question is worthy of further study.

3.10 The Doubling Time of Sea-Level Intensification and Separation Distance

In dynamic and synoptic meteorology the rate of intensification of unstable wavesis important. Phillips (1954) and others have published values of the theoretical doubling time of the growth of instability waves, and Petterssen (1956)[2] has given typical doubling

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[1]This question will be examined further in Chapter 5.

[2]See page 293.

times for atmospheric disturbances of various scales.

In the present study, the mean doubling time of moderate to intense lee cyclones was evaluated. The doubling time was obtained from plots of intensity (∇^2 p) versus time as shown in Fig. 10. However, the doubling time may be also deduced from computations of the vorticity of sea-level cyclones, as was done by Hage (1957), and Fawcett and Saylor (1965).

The weak disturbances of Types C and D will not be considered here since their intensities were uncertain. However, the moderate local cyclones of Type B were found to intensify normally six hours after t_o. But only one cyclone (case 131) actually doubled in intensity while the others developed considerably, but did not quite double in intensity. In addition, these cyclones were smaller, and the weather phenomena were less wide-spread and dramatic. One exception, case 94 of Type B, appeared to be associated with an upper wave for part of the time.

Little more than half of all Type A cyclones doubled their intensity during their life spans, and the remainder attained nearly double intensification. Most of these cyclones reached their maximum intensities within 6 to 12 hours, and only rarely within 18 to 24 hours. The results agree with those of Petterssen, who found doubling times of 6 hours for intense synoptic motion systems, and 24 hours for medium synoptic systems.

It is also of interest to consider the time lag between the time of cyclogenesis and of maximum intensity of the cyclone. On the whole, the lag was found to be about 36 to 48 hours, but some cyclones took as little as 18 to 24 and as long as 52 to 72 hours to reach

maximum intensity. It was also observed that a majority of cyclones began to intensify rapidly, with the approach of upper troughs, some 12 to 18 hours after initial formation. This indicates possibly that, influenced by orographic processes, lee cyclones intensify on the average about 12 hours earlier than "normal" non-orographic cyclones.

A rough estimate of the average time interval between the first and second intensification, t_1 and t_2, for the development Type A_2 cyclones may be of interest. The average time for a typical low was about 28 hours, corresponding to travel distances of 600 to 1,200 km, depending on a system's speed, and with large variations for individual cases.

Returning to the question of separation distance, the curves plotted in Fig. 23 depict the crucial development phase of two types of cyclone. The top two curves in the figure are for typical cases of lee development of Type B and the lower curves are for developments of Type A_1 and A_2. The horizontal separation distance between an upper trough and a lee cyclone was estimated from the 500-mb and surface charts.

In the case of most Type B cyclones, the separation distance is in excess of one thousand kilometers, which all but assures that the vorticity maximum is also well upstream and far from the site of the nascent surface low. Considered in another way, in the sense of a corollary, it *appears that while Petterssen's hypothesis does play a central role in lee cyclone intensification, it does not in general account for lee cyclone initiation of Type B cyclones.*

As seen in the lower part of Fig. 23, the separation distances for typical intense developments were comparatively large during the

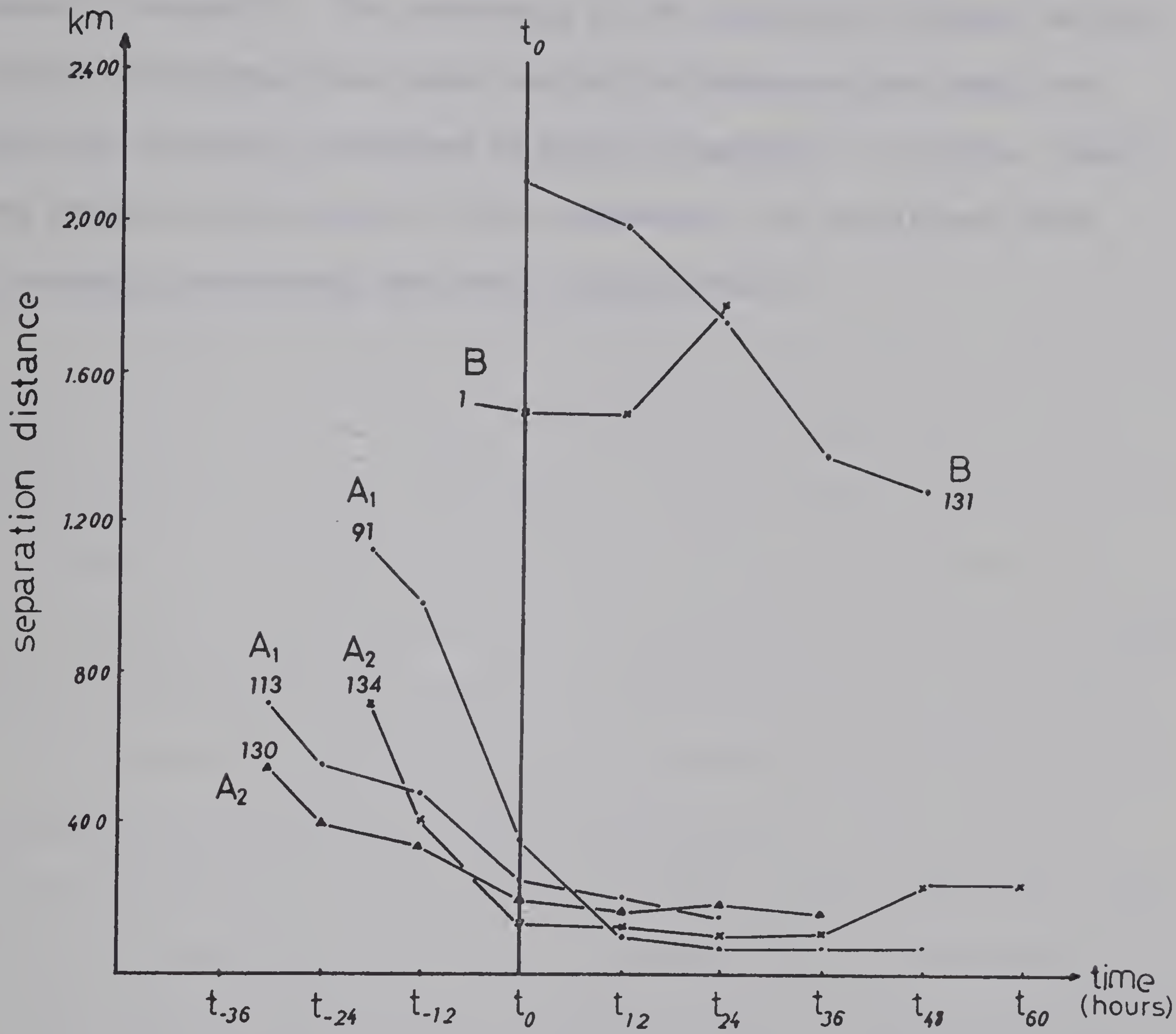


Fig. 23. Typical separation distances between upper cold lows and associated sea-level cyclones. The upper two curves show Type B development, and the lower curves are for Type A_1 and A_2 development. Successive 12-hourly map time is plotted along the abscissa, and separation distance along the ordinate. The t_o ordinate marks the time of onset of sudden intensification of Type A_1 and B cyclones; in the two cases of Type A_2 development, this ordinate marks the onset of the second intensification (t_2). Block numbers indicate specific case histories.

early stages, but they decreased sharply in the period t_{-12} to t_o, and in most cases continued to decrease slowly thereafter. Thus the surface and 500-mb systems tend to be in close proximity when cyclones begin to intensify. The shortening of the separation distance in lee cyclone developments was noted earlier by Petterssen and Hage, and has been variously considered as being "orographic" in nature. Awaiting further investigation of this phenomenon, the provisional term "orographic shortening" may not be inappropriate.

CHAPTER IV

SOME CASE STUDIES OF LEE CYCLOGENESIS WITH REFERENCE TO OROGRAPHIC MOTIONS

4.1 Typical Cases of Development Type A_2

Since the energetic lee cyclones of Type A_1 and A_2 differ mainly in the characteristics of their intensification, only cases of Type A_2 will be discussed. Figs. 24, 25 and 26 were prepared to illustrate the development of two Mackenzie cyclones of Type A_2, cases 99 and 100. Lee cyclones of this type which have originated in the other two principal source regions have been observed to develop in much the same manner.

An intense surface low has drifted into Alaska and is in the process of dissipating, as can be seen in the upper left of the two surface charts in Fig. 24. In the mean time, a weak cyclone has formed in the Southern Yukon, and a weakly diffluent 500-mb pattern containing a vorticity minimum has become established over the Mackenzie Mountains. This diffluent flow, essentially perpendicular to the mountains, showed up also on the previous upper air map (not reproduced here). A vorticity maximum is situated over Southeastern Alaska at this stage. The two TROWALs shown on the surface map were weak and did not affect the flow pattern on the 500-mb level.

In the next 6 hours (by 0600Z; chart not shown), a cyclone (99) formed with a closed isobar in the lee; the low began to intensify thereafter, while weak vorticity advection was evident over it. Mean-

while, the original surface and upper cold lows remained over south-western Alaska. The separation distance between cyclone (99) and the upper cold low was large. The 1200Z, 2 October sea-level chart, Fig. 24-b shows that the low has intensified and moved along with the upper flow. During this time, a weak vorticity maximum, produced and maintained by the horizontal shear flow aloft, gave support to the sea-level cyclone. Despite the absence of an upper trough, the surface low attained its maximum intensity by 1200Z, 2 October, with a doubling time of 6 hours. This first intensification seems to be mainly due to the shear vorticity of the jet stream in the converging upper contour pattern over northeastern Alberta.

Attention should be paid again to the diffluent pattern over the northern half of the Mackenzie Divide as shown on the 1200Z 500-mb map of 2 October. There is also some diffluence in the non-normal flow crossing the Northern B.C. Range, but the perpendicular flow is again maintained over the Mackenzie Mountains.

Turning to Fig. 25, it will be noticed that a new lee cyclone (100) has been already initiated on a front under the diffluent flow normal to the Divide, while a vorticity minimum at the 500-mb level was located just to the east of the nascent cyclone. Positive vorticity advection has set in, and the cold trough shown on the 500-mb maps has deepened suddenly just east of the Mackenzie Mountains, while the surface low has moved to Northern Alberta. This cyclone began to intensify from 1800Z onward with the development of the upper trough, and finally attained maximum intensity at 0000Z, 4 October, with a doubling time of 6 hours. The former lee cyclone (99) is already in the process of filling since it acquired its maximum intensity, as

seen on surface charts (a) and (b).

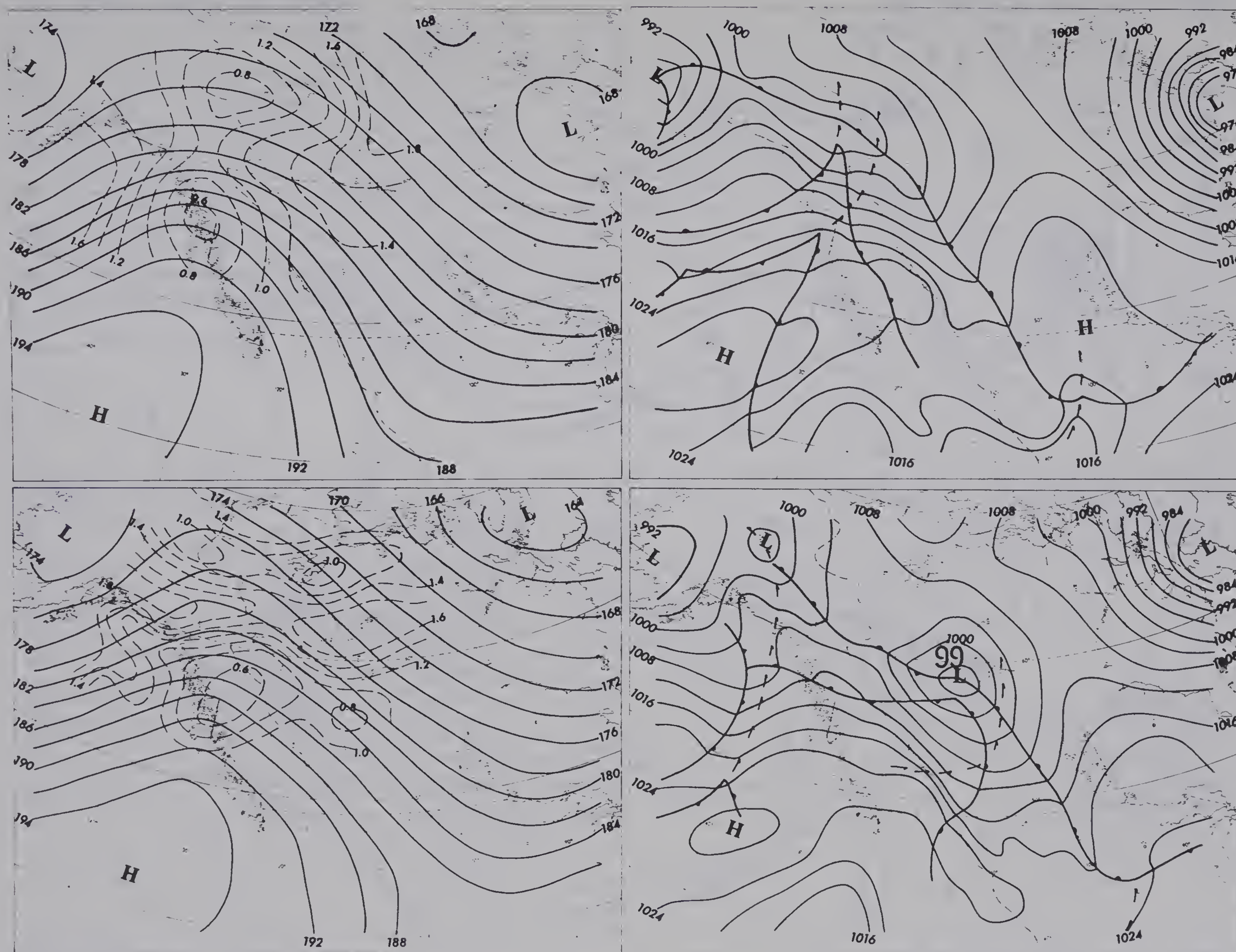
Referring to Fig. 26-a, it will be seen that the cyclone (100) has now intensified markedly under strong positive vorticity advection aloft. At the same time a broad, diffluent upper pattern lies again normal to the Divide of the Mackenzie Mountains. This pattern is associated with the nascent wave on the 0000Z, 4 October surface chart, and initiated the new cyclone (102) shown on the 1200Z chart (b). This cyclogenesis clearly did not involve vorticity advection. Meanwhile, cyclone (99) has moved to James Bay, filling on the way. Cyclone (100), on the other hand, acquired its second maximum of intensity at about 0600Z, 5 October (chart not shown) with a doubling time of 12 hours. By then, cyclone (99) had all but disappeared, having been "swept up" by the more intense and upper-trough supported cyclone (100).

Other synoptic features in this series worthy of note are the dominant warm anticyclone over Western North America and the adjoining Pacific, and the outbreaks of arctic air into Central Canada. Such circulation régimes usually lead to the establishment of a low-level frontal system in the lee of the Divide. On such fronts, cyclogenesis in the sense of the classical polar-front theory appears possible. The existence and significance of quasi-stationary lee fronts has been commented on by Hage (1957) and Newton (1958; see Reiter, 1963).

Returning to the original intense sea-level cyclone over Alaska, and the associated upper cold trough, it will be noted that this system was in a state of continuing cyclolysis. But directly and indirectly this system contributed to the generation of three cyclones in the lee of the Mackenzie Mountains. The first two of these cyclones amalgamated again in the vicinity of James Bay after crossing

(1) 500-mb chart for 0000Z

(a) Surface chart for 0000Z



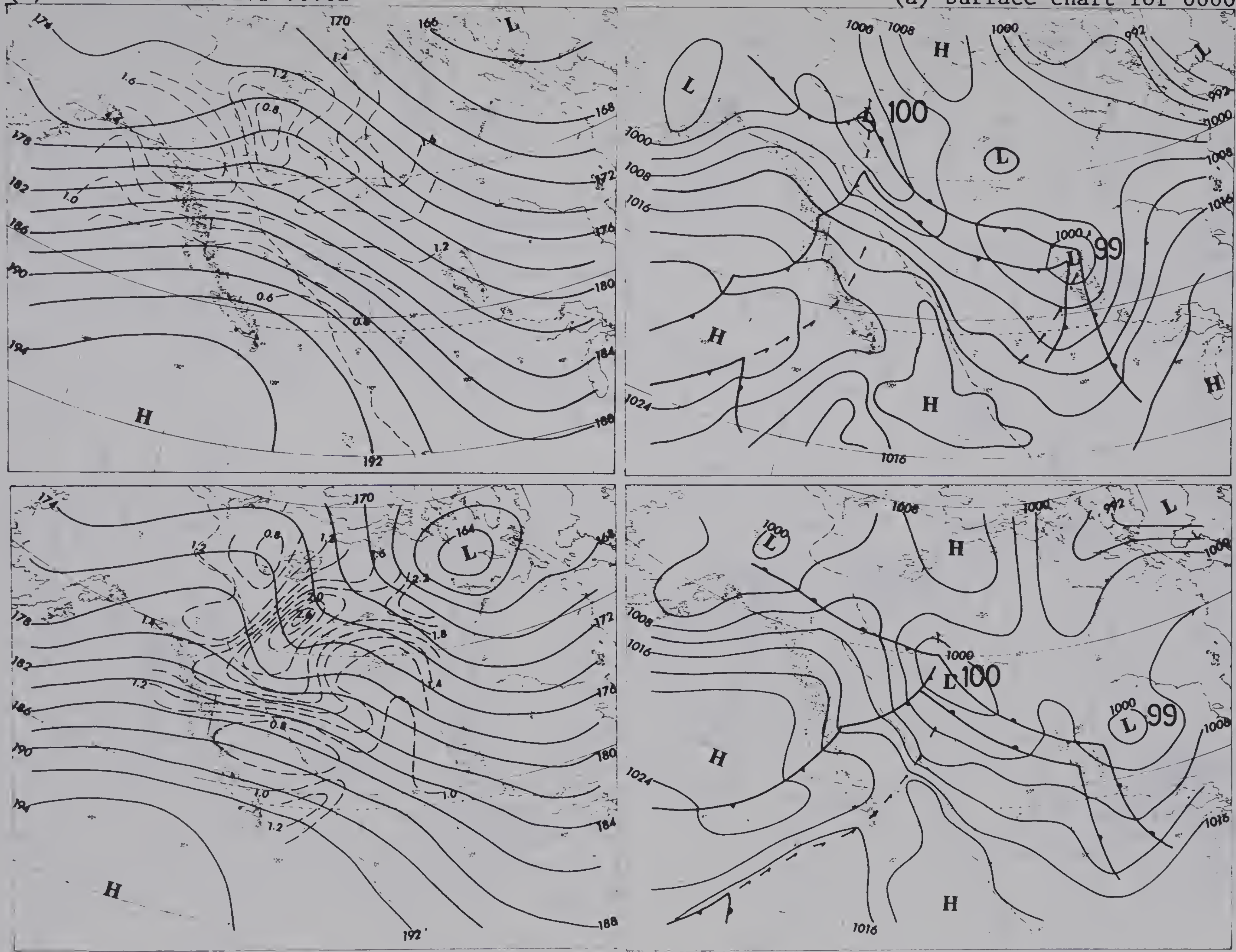
(2) 500-mb chart for 1200Z

(b) Surface chart for 1200Z

Fig. 24. Charts for 2 October 1958.
Contours at 200 foot intervals. Isopleths of absolute vorticity in 10^{-4} sec^{-1}.

(1) 500-mb chart for 0000Z

(a) Surface chart for 0000Z



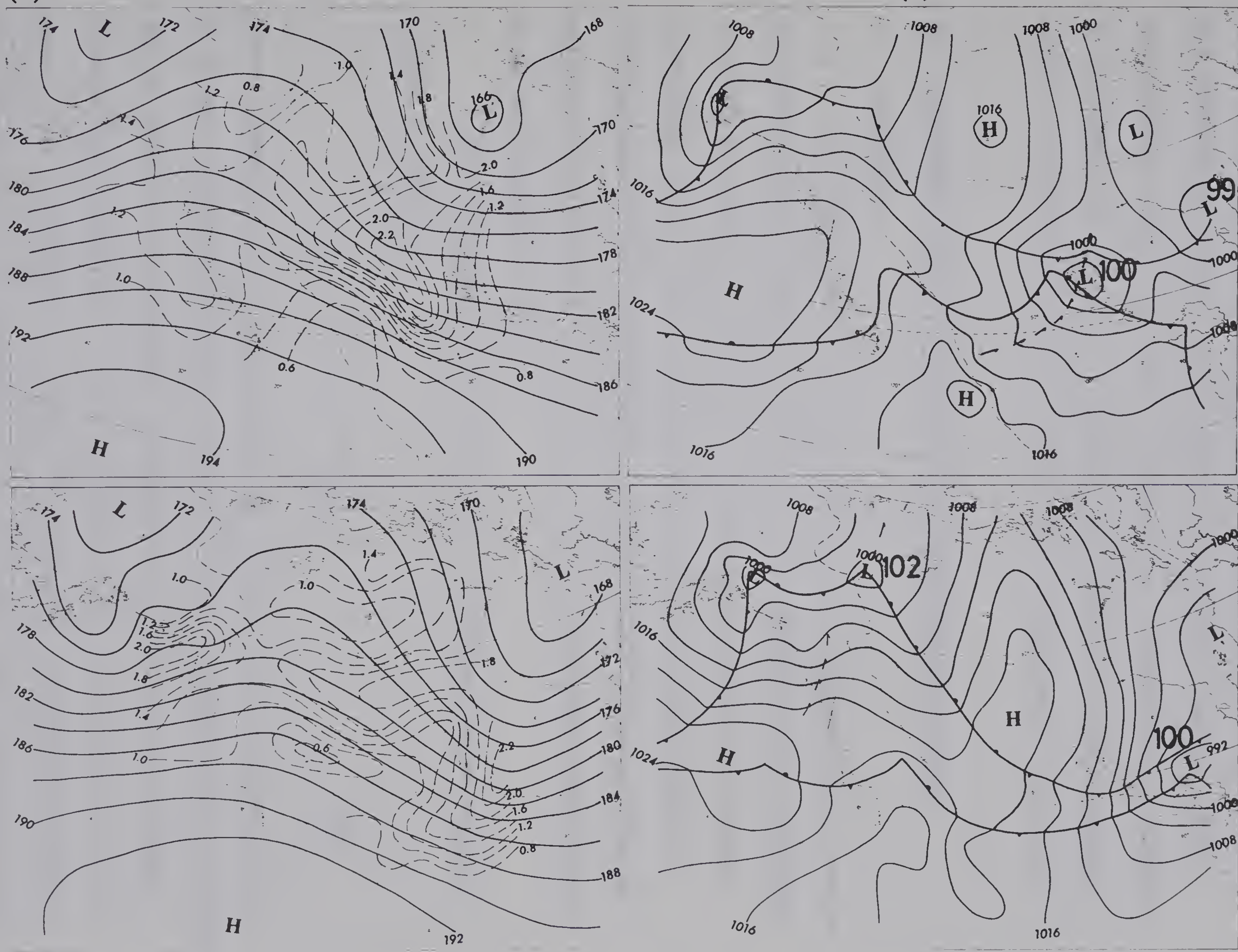
(2) 500-mb chart for 1200Z

(b) Surface chart for 1200Z

Fig. 25. Charts for 3 October 1958. See legend for Fig. 24.

(1) 500-mb chart for 0000Z

(a) Surface chart for 0000Z



(2) 500-mb chart for 1200Z

(b) Surface chart for 1200Z

Fig. 26. Charts for 4 October 1958. See legend for Fig. 24.

half the continent as separate entities. Similar west to east moving series of cyclones, originating from a parent low over the Pacific, are produced in the lee of the other two principal ranges. These developments may be the result of some kind of instability pulsation of the upper air motions. Another similar form of lee development, a north to south moving series of lows, will be discussed in section 4.3.

4.2 Orographic Motions Associated with Developments of Type A_1 and A_2

An attempt was made to account for the orographically induced low level vertical motions in the lee of the Canadian Rocky Mountains. In order to compute the motions, the smoothed height profile shown previously (Fig. 8) was converted into a pressure field using the International Standard Atmosphere. The converted pressure field was again analyzed by means of isobars, but it is not shown here because it is very similar to Fig. 8.

The following equation was used to evaluate forced vertical motions in the lee:

$$\omega_o = \vec{V} \cdot \nabla p \simeq V\left(\frac{\Delta p}{\Delta s}\right) \tag{15}$$

where V measures the actual horizontal wind at a fixed sloping surface Δp is the pressure difference between two points on the sloping surface; and Δs measures the horizontal distance in the direction of the wind between the same points. For convenience, only two values of Δs were used in the evaluation: A length of one degree of latitude was used for steep slopes, and two degrees for relatively gently-sloping terrain.

Prior to the evaluation, a problem was encountered regarding the use of representative winds, since the lee slope descends from about 700-mb at the Divide to sea-level at Hudson Bay. Most of the mountain wave theories discussed in section 2.5 consider the atmospheric level which intersects a mountain range at the half-width distance[1] to be of particular interest, in that it is the level at which strong vertical stretching is most likely to occur. In the present case, this level corresponds closely to the 850-mb level. Hence the winds reported on 850-mb charts were used in the evaluation, although it is likely that this procedure will lead to under- and over-estimates for levels higher and lower, respectively, than the 850-mb surface. Proceeding in a manner similar to that employed in constructing the smoothed topography in section 3.2, orographic vertical velocities were estimated at every 2.5 degrees of latitude and longitude in the lee region, and analyzed in terms of vertical wind isotachs, at intervals of 1×10^{-3} mb/sec.

The series of four charts in Fig. 27 illustrates the development of the Mackenzie low of case 100. These charts should be used in conjunction with Figs. 25 and 26. *With moderate descent in the lee of the Mackenzie Mountains (see chart 1), weak cyclogenesis leading to a closed isobar about the centre occurred at sea-level, in close proximity to isotach 3, and below a diffluent 500-mb pattern.* Vorticity advection was not vigorous in the initial stages of this cyclogenesis.

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[1]The half-width distance is defined as the projected horizontal distance measured from the crest of a mountain to a point on its slope one half as high as the crest.

Fig. 27-1 shows also *much larger descending motions* in the lee of the Northern B.C. and Southwestern Alberta Ranges. However, *without adequate vorticity advection, and the absence of a quasi-normal diffluent pattern aloft, cyclogenesis did not take place in the lee of these two ranges. This suggests that orographic down slope motions per se do not play a leading role in the initial formation of lee cyclones.* This is somewhat at variance with a discussion of lee slope motion by Palmén and Newton (1969)[1] but lends indirect support to Newton (1956) who found that frictional effects are important factors in the early stage of cyclogenesis.

From Fig. 27-2, it will be noticed that descending motions in the lee of the Mackenzie Mountains have increased slightly in magnitude as well as in area. The surface cyclone attained its first maximum intensity at 0000Z, 4 October, the time of the third map in Fig. 27, when positive vorticity advection with an upper trough (see Fig. 26-a) was at a maximum.

Another interesting aspect, shown in Fig. 27-3, is that an extended area of slow ascending motions has developed over Alberta, adjoining the region of descent in the immediate lee of the Southwestern Alberta Range. As pointed out in section 3.2, the topography of the lee slope is complex and the pattern of vertical motions is likely to be complicated, with alternate regions of ascent and descent. It is not unreasonable to expect these motions to contribute significantly to development, in view of the tendency equation of Margules and Bjerknes, and the Petterssen development equation, which include vertical motion terms.

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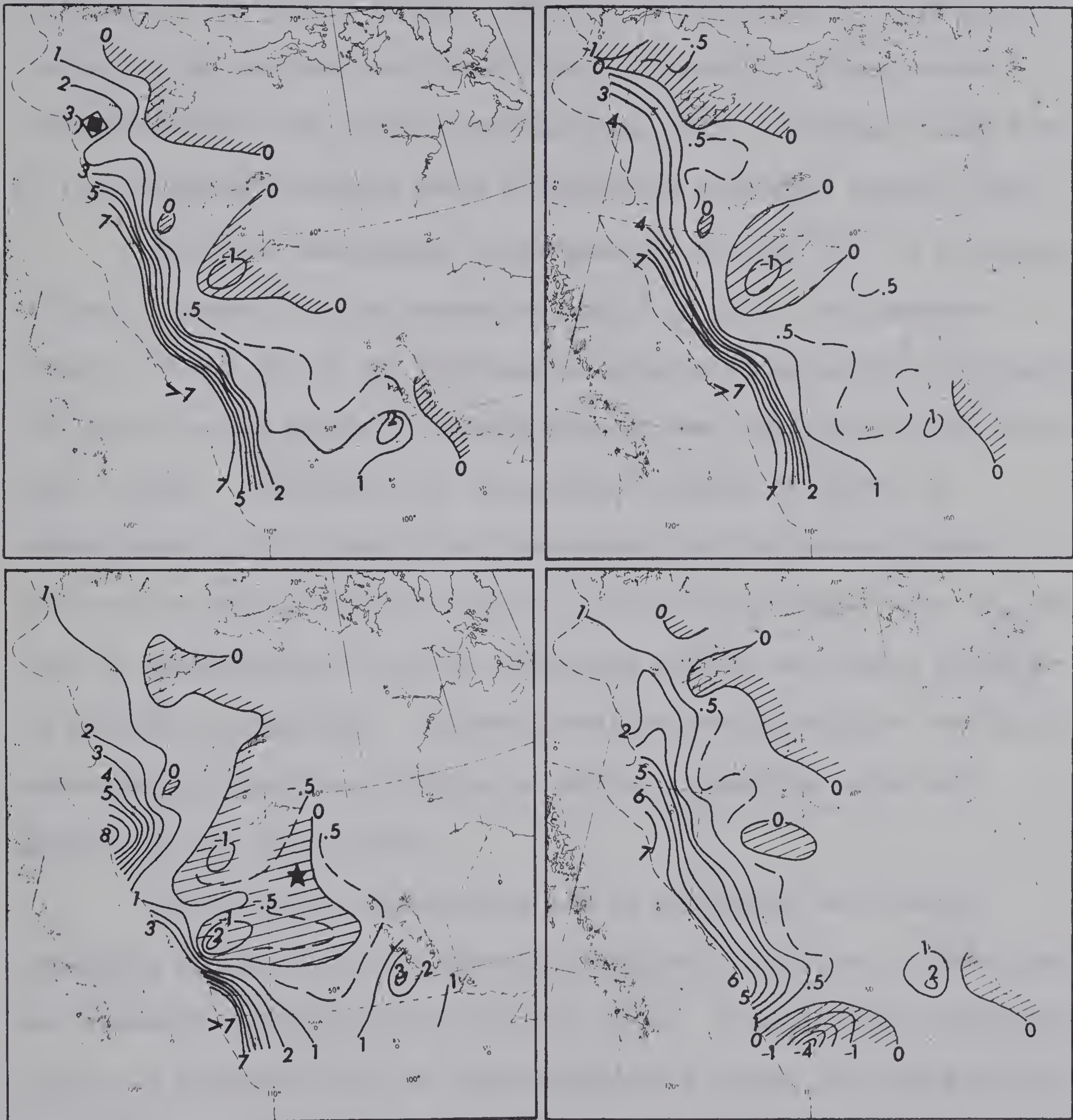
[1]This problem will be discussed further in Chapter 5.

(1) 0000Z, 3 October 1958 (2) 1200Z, 3 October 1958



(3) 0000Z, 4 October 1958 (4) 1200Z, 4 October 1958

Fig. 27. Orographic vertical velocities (ω_0 in units of 10^{-3} mb sec^{-1}) associated with the cyclone (100) of development Type A_2. The symbol ◆ in the lee of the Mackenzie Mountains of the first chart marks the position of the initial appearance of the cyclone; ★ in the third map indicates the position of the first intensification of the cyclone.

In Fig. 27-4, the extended area of slow ascent in Alberta has been replaced by strong downslope motion, now that the cyclone has moved out of the source region. There is no cyclogenesis occurring in the lee of the Southwestern Alberta and Northern B.C. Ranges under these conditions, but another cyclone (case 102) is forming in the lee of the Mackenzie Mountains under a pronounced diffluent pattern aloft.

A similar development is illustrated in Fig. 28, for a cyclone of Type A_1 (case 6). The general synopsis[1] is that a cyclogenesis occurred in the lee of the Southwestern Alberta Range under a pronounced diffluent flow at 1800Z, 12 January without the close association of an upper trough. The surface low intensified rapidly by 0600Z, 13 January when a cold trough aloft approached the Continental Divide and moved to within 500 km of the low. It will be noticed from Fig. 28 that the orographically induced descending motions were quite large at the time of cyclogenesis. Moreover, small ascending motions were also maintained over Northern Alberta as the low intensified with the approach of the upper trough.

The foregoing observations may be summarized as follows: Descending motion leads to vertical stretching of the air columns, and the production of vorticity on the lee slope. If a diffluent mid-level pattern of divergent flow is superimposed on a region of orographically induced descending motion, lee cyclogenesis is likely to ensue.

It appears plausible to conclude that *the initiation of lee*

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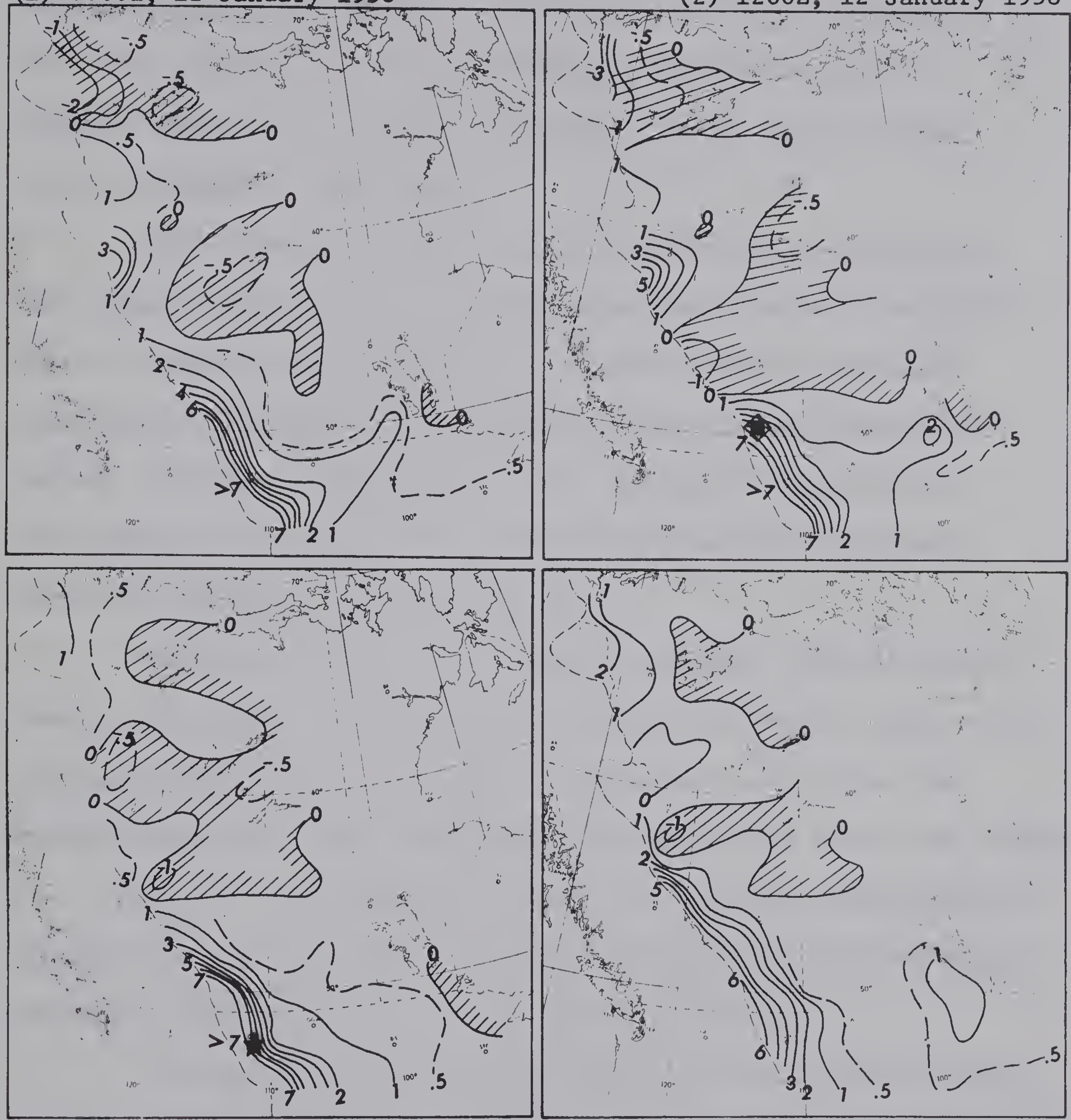
[1]The synoptic charts are not reproduced here.

(1) 0000Z, 12 January 1958 (2) 1200Z, 12 January 1958



(3) 0000Z, 13 January 1958 (4) 1200Z, 13 January 1958

Fig. 28. Orographic vertical velocities (ω_o in unit of 10^{-3} mb sec^{-1}) associated with a cyclone development of Type A_1 (case 6). The symbol ◈ in the lee of the Southwestern Alberta Range on the second chart marks the position of the initial appearance of the cyclone at 1800Z, 12 January; ★ in the third map indicates the position of the intensification of the cyclone at 0600Z, 13 January.

cyclogenesis occurs in this case primarily in response to the influence of upper level, quasi-normal, diffluent velocity fields and that of orographic descending motions. Maintenance of flow continuity may in turn force regional low level convergence, and produce cells of vertical motion in the middle troposphere, in the sense of Dines' cellular scheme of compensation.

The structure of Dines' cellular patterns of compensation will depend greatly on the location of the levels of non-divergence. There is little in the literature to indicate what the shape and location of the levels of non-divergence might be over mountainous terrain. However, according to Reinelt (personal communication) there should be at least two levels over the Canadian Cordillera where the divergence is effectively zero.

The first "level", *the orographic surface of non-divergence,* should exist close to the terrain, and conform in general shape to the mountain profile, for it is near the orographic barrier that the vertical velocities will reach maximum values for the first time, making $D = - \partial\omega/\partial p = 0$. The maximum elevation of this level should coincide closely with the height of the mountains along the Continental Divide, and should therefore be not much in excess of 700 mb.

A second level, *the synoptic level of non-divergence,* should occur over the Divide at about 350 mb, lowering thence to the "standard" height of about 600 mb to the windward and leeward, i.e., well out over the Pacific and the Central Prairies. This is the level at which the synoptic-scale vertical motions are likely to attain maximum values.

The location of the two main levels of non-divergence largely governed the choice of the 500-mb chart as being the most appropriate to show the flow pattern aloft, for it seemed unlikely that the 500-mb level would coincide with a level of non-divergence in the region under study.

Accepting the high incidence of cyclonic activity in the lee of the Rockies as an established fact, it would seem reasonable to postulate the existence of a preferred, i.e., orographically determined pattern of cellular compensation which, when incomplete, strongly enhances lee cyclogenesis. In this view, the phenomenon of lee cyclone initiation falls naturally within the compass of the development theories of Margules, Bjerknes, Dines and Scherhag. The sudden intensification of an existing cyclone seems to depend, on the other hand, largely on positive vorticity advection aloft - an observed fact in basic accord with Petterssen's development hypothesis.

4.3 North-South Series of Multiple Lee Cyclogenesis

Reference was made in the section 2.1 to the possibility of the development of a series of cyclones at various points in a lee trough, in situations when a broad zone of strong cross winds is present over an extensive range of mountains. In the course of this study at least ten synoptic situations were encountered which generated two or three cyclones in different places in the lee of the Divide. In general, it was found that such multiple cyclogenesis was associated with a parent cyclone and an upper trough in the Pacific. Hage (1957) has observed the similar development of 2-3 surface cyclones with only

one upper cold low, and concluded that secondary lows tend to form south of the primary one.

The site of cyclogenesis of the more vigorous members of a series was initially determined by the position of the diffluent cross-flow with respect to the Divide, and by the relative strength of the wind field aloft. Also, the time and place of formation of the lesser members in a series were usually determined by such conditions, and most tended to be steered by, or moved along with, the upper flow. Furthermore, it was observed that in general only one of a series underwent moderate to intense development and, moreover, these cyclones often amalgamated again at some later time.

The early stages of multiple cyclogenesis of this kind seem to be strongly dependent on orography and the position of the diffluent pattern aloft. If both a jet stream and a diffluent pattern are present within a broad band of quasi-normal, cross-barrier flow some ten or more degrees of latitude in width, then multiple cyclogenesis will likely occur. Simultaneous formation of two or more lee cyclones is rare; usually, the members of a family form singly, and at distinct times. The site of cyclogenesis of the first member of a family is usually determined by the location of the principal zone of diffluent cross-barrier flow, and subsequent members all tend to form in places where secondary diffluent patterns develop quasi-normal to the Divide.

In addition, *there is clear evidence that, if similar diffluent conditions prevail over the Northern B.C. Range and the Southwestern Alberta Range at the same time, the Alberta cyclone grows much more intense.* This may be the result of the different strength of the orographic vertical motions. If the wind field is favourable over the

northern mountains, then a secondary low will develop to the north of the primary Alberta low. Since the 500-mb diffluent patterns are usually well defined and readily identified, they are useful prognostic indicators of impending lee cyclogenesis.

4.4 The Development of a Type B Cyclone and the Orographic Motions Associated with the Development of a Lee Trough

In sections 4.1 and 4.2 trajectory Type A cyclone developments were discussed. A moderate local Type B cyclone will be described here, which was initiated in the lee of the Coast Range, and subsequently was absorbed into a lee trough that formed in the lee of the Rocky Mountains.

From Fig. 29 it will be seen that a cyclone of moderate size and development is moving into the Gulf of Alaska, and toward some ill-defined frontal systems in the Yukon. A pronounced upper ridge along the West Coast is slowly moving inland. By 1200Z, 27 November, the upper ridge is over northern B.C. and the Yukon with diffluence and a vorticity minimum over the lee of the northern Coast Range and the Wrangell Mountains. The corresponding 1200Z surface map shows a weak cyclone with one closed isobar (case 131), which developed in the lee of the Northern B.C. Coast Range, slightly to the east of a pronounced diffluent 500-mb contour pattern. This cyclone was initiated with a separation distance of about 2,000 km. As shown in Figs. 30 and 31, this cyclone centre was short-lived, and had degenerated into a trough by 0000Z, 29 November. No positive vorticity advection nor minor upper trough could be identified or associated with this local development.

The diffluent upper air pattern ceased to progress eastward, and the surface low failed to grow when the upper ridge had become stationary and tended to retrograde (see Fig. 31). The process of cyclolysis was all but completed by 0000Z, 29 November, leaving a residual lee trough which, however, intensified subsequently with the approach of the rapidly deepening low on the Pacific. During this period low-level descending motions were taking place in the lee of the Rocky Mountains, as shown on all four maps in Fig. 32. At the time of cyclogenesis, 1200Z, 27 November, orographic ascending motions prevailed to the east of the lee cyclone (see charts 2 and 3 in Fig. 32). *Though strong orographic motions were present in the lee of the Southwestern Alberta Range, the cyclone appeared first in the lee of the B.C. Coast Range where the diffluent flow aloft was almost normal to the ridge of the mountains.* However, some 24 hours later, and despite the sharp increase in the orographic vertical motions over the Northern B.C. Range, the low level cyclone continued to decay, apparently due to the lack of adequate upper air divergence.

As just mentioned, there were strong downward motions in the lee of the Southwestern Alberta Range, but cyclogenesis occurred only in Northern B.C. This lack of development in Alberta may be ascribed to the presence of a confluent upper pattern over the Southwestern Alberta Range, and the ensuing marked convergence in the lower and middle troposphere. It seems clear, therefore, that *strong orographic downslope motions alone do not determine the initial formation of lee cyclones nor their subsequent deepening, but such flow régimes can produce and maintain lee pressure troughs.* A good example of such a lee trough is shown in Fig. 31 of this synoptic series.

(1) 500-mb chart 0000Z

(a) Surface chart for 0000Z

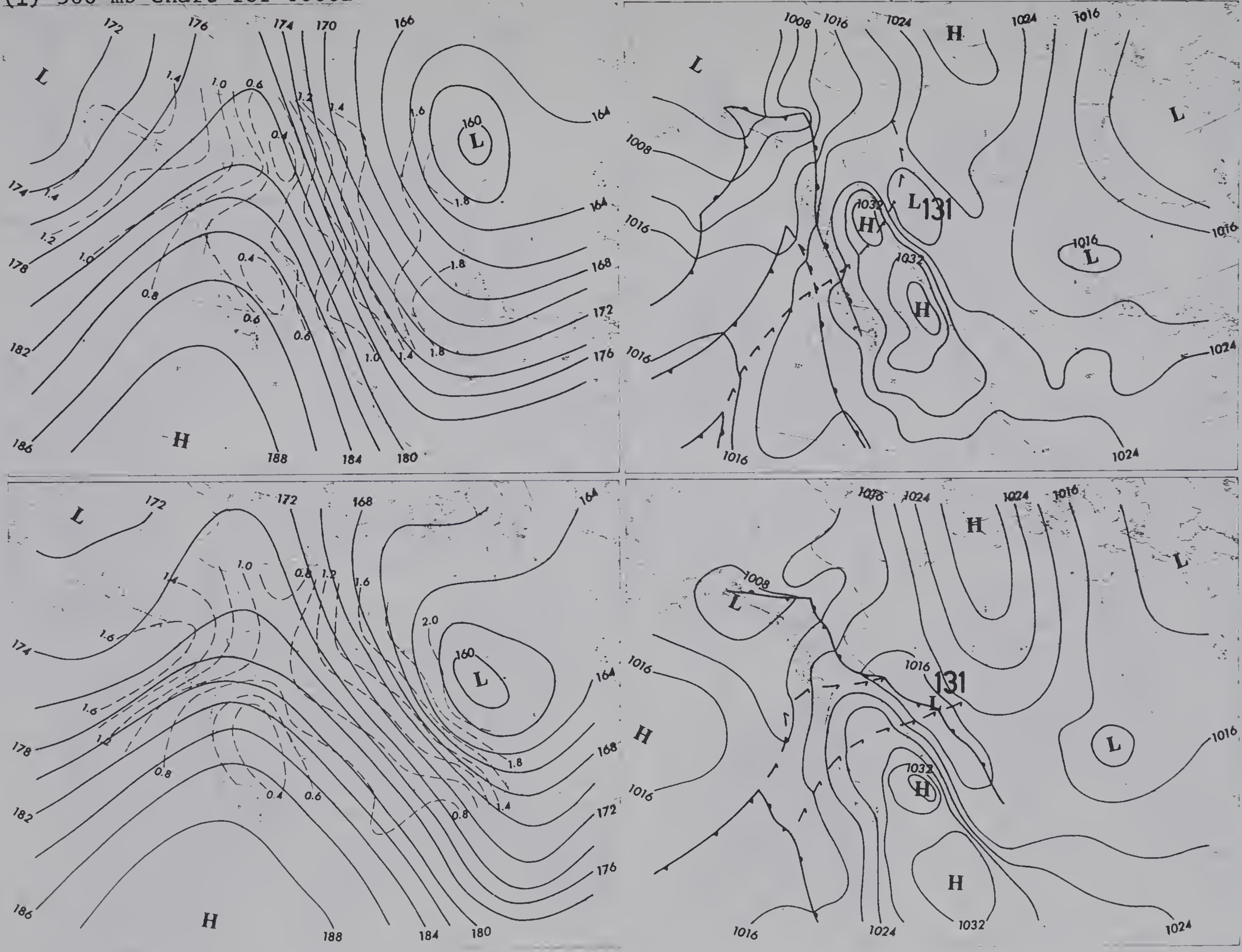
(2) 500-mb chart for 1200Z

(b) Surface chart for 1200Z

Fig. 29. Charts for 27 November 1958. See legend for Fig. 24.

(1) 500-mb chart for 0000Z

(a) Surface chart for 0000Z



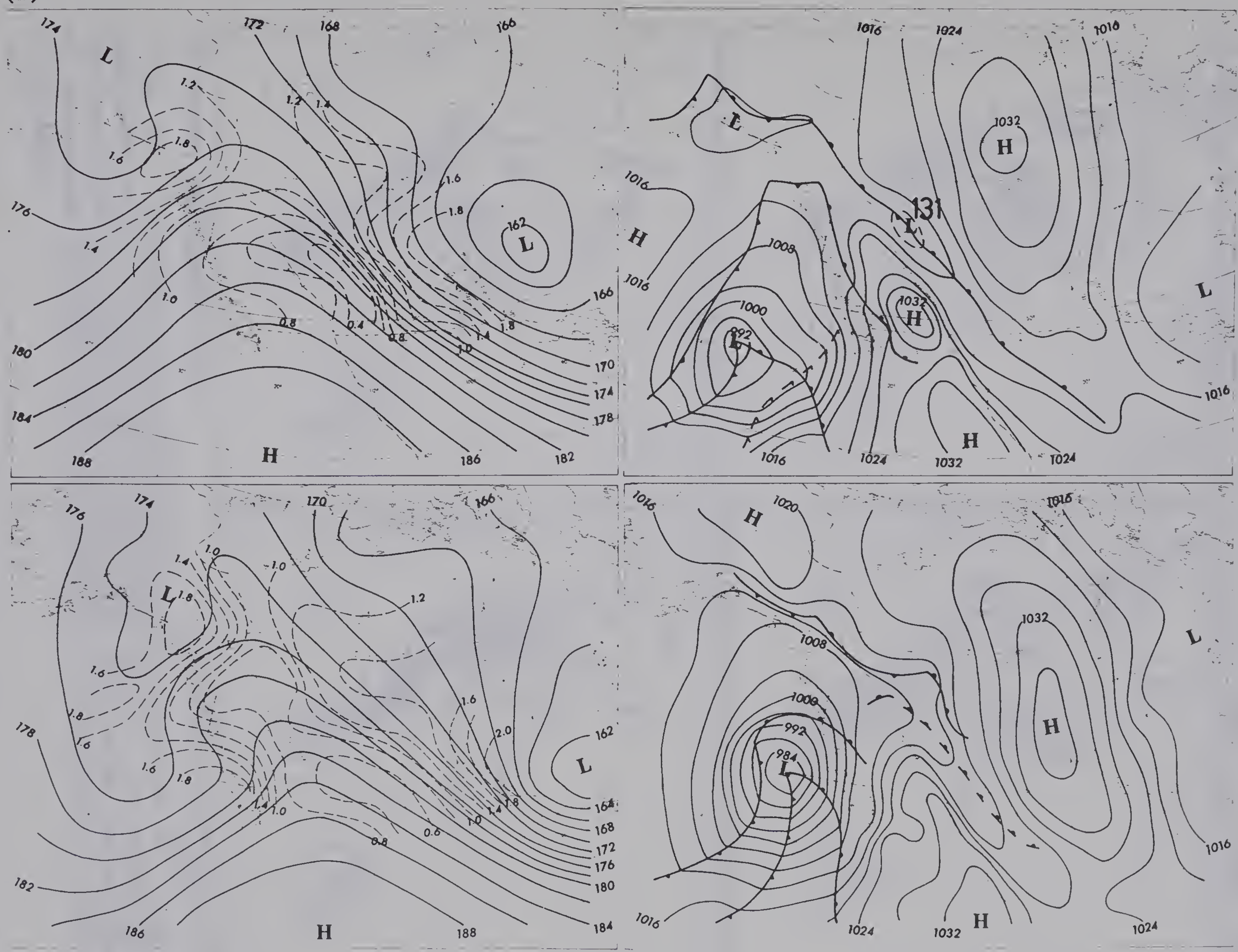
(2) 500-mb chart for 1200Z

(b) Surface chart for 1200Z

Fig. 30. Charts for 28 November 1958. See legend for Fig. 24.

(1) 500-mb chart for 0000Z

(a) Surface chart for 0000Z

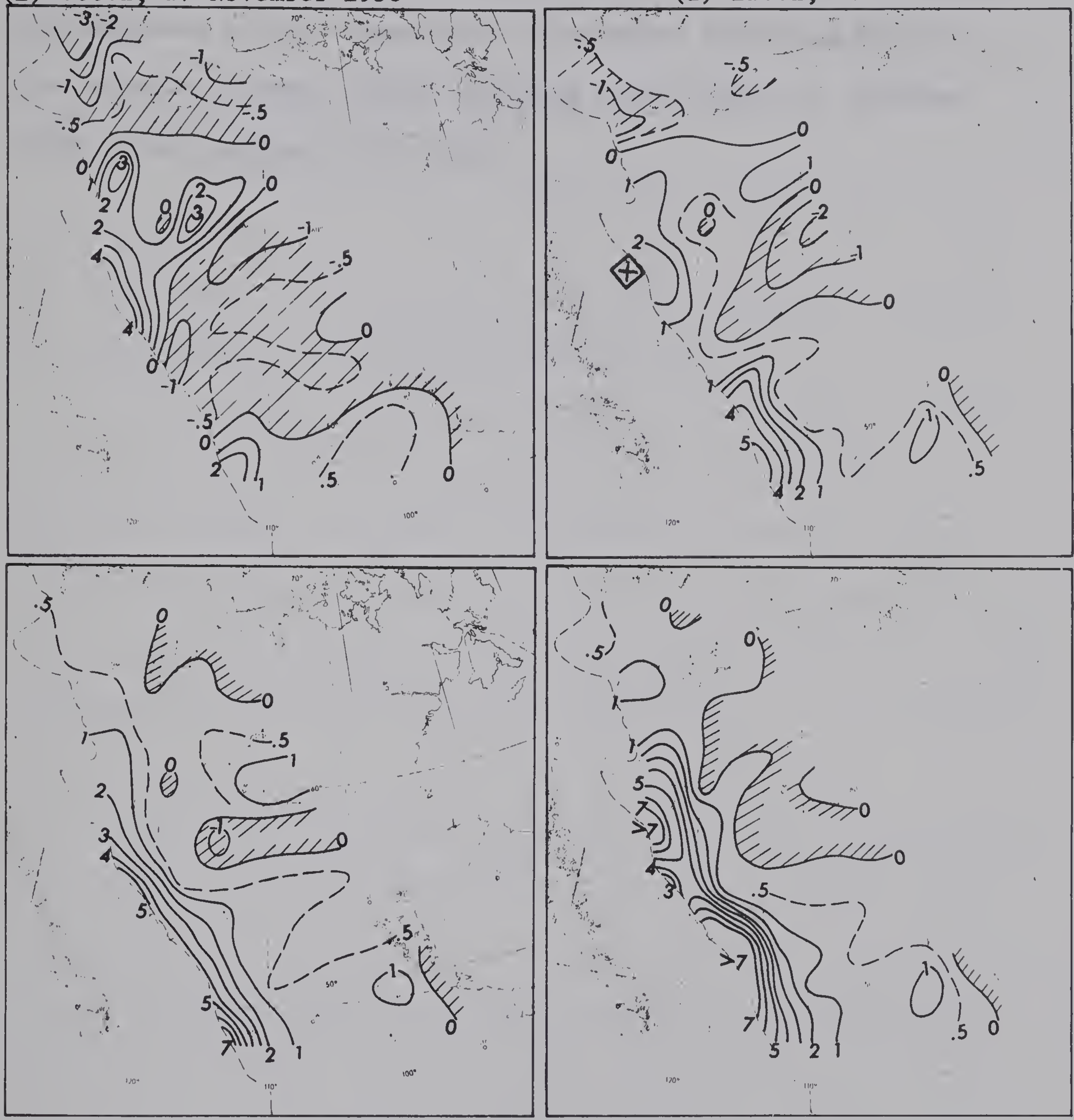


(2) 500-mb chart for 1200Z

(b) Surface chart for 1200Z

Fig. 31. Charts for 29 November 1958. See legend for Fig. 24.

(1) 0000Z, 27 November 1958 (2) 1200Z, 27 November 1958



(3) 0000Z, 28 November 1958 (4) 1200Z, 28 November 1958

Fig. 32. Orographic vertical velocities (ω_0 in unit of 10^{-3} mb sec^{-1}) associated with the cyclone (131) development Type B. The symbol $\oplus$ in the lee of the B.C. Coast Range of the second chart marks the position of the initial appearance and intensification of the cyclone at 1200Z, 27 November.

At any rate it appears certain that low-level convergence in an air column cannot alone account fully for cyclogenesis without the imbalance in the compensation of vergence fields implicit in the theories of Dines (1919), Scherhag (1934, 1937), J. Bjerknes (1937, 1944) and Sutcliffe (1939).

CHAPTER V

SUMMARY AND CONCLUSIONS WITH SOME COMMENTS AND SPECULATIONS ON SYNOPTIC-SCALE BARRIER FLOW

5.1 Some Comments on Orographic Effects, Lee Cyclogenesis and the Intensification of Diffluent Upper Patterns

Many authors, most notably Queney (1948), Hess and Wagner (1948), Scorer (1949), and Petterssen (1956) referred to earlier in Chapter 2, showed or suggested that vertical shrinking of the air column must occur over the windward side, and vertical stretching in the lee of mountain ranges. These processes in turn lead to, or are associated with vorticity destruction and generation in cross-barrier flow.

Schallert (1962) felt that "such stretching in appreciable amounts appears to be present only under special conditions, and the mechanisms that bring about such vertical stretching in the lee are not known". Radinović (1965) considered the effects of the Alps on the 1000-500 mb thickness deformation, and cyclogenesis in the Gulf of Genoa. Using a baroclinic model, he showed that forecasts of the 500-mb contours were substantially improved when orographic deformation was included in the model.

As pointed out in earlier chapters, the isobaric vergences are, of course, known to be very important in the development of weather systems. It is also well known that anticyclones build in intensity

over the windward side while cyclones decay. However, it seems that the horizontally divergent motions over mountains and the part played by them in lee cyclogenesis were not adequately emphasized in some of the earlier work, though a great many studies have been done on the sudden intensification of lee cyclones.

It is clear that forced ascending motions are produced by the incoming winds on the windward side of mountain barriers. These motions will be examined qualitatively with reference to the simple barrier profile shown in Fig. 33. The profile in the upper half of the figure represents a mountain range several hundreds of kilometers in extent in a north-south direction, with west to east symmetry about the 3 km summit and a half-width of 100 km; steady, strong and generally westerly winds prevail over the barrier under conditions of frictionless flow. According to the various theories, the forced vertical motions will produce vertical shrinking and stretching of the air columns passing over the windward and leeward sides, respectively. If the conservation theorem of potential absolute vorticity, and the equation of continuity in the form $D_p = -\frac{\partial\omega}{\partial p}$ are considered together, *one should find horizontal divergence at all isobaric levels over the windward slope where* $\partial\omega/\partial p$ *is negative, even with contour flow patterns which are originally parallel or slightly confluent.* Moreover, *since the trajectories of air particles on the windward side acquire anticyclonic*[1] *curvatures, and the Coriolis parameter increases with*

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[1]See Bjerknes et al. (1933) and Queney (1948). Detailed analysis shows that the flow will first be deflected northward with cyclonic curvature some distance upstream of the barrier (as shown in Fig. 33-b) before acquiring anticyclonic curvature. According to C. Charette (1972, personal communication), this problem was recently also the subject of correspondence between H. P. Wilson and B. W. Boville.

latitude, the contour pattern of a broad current will tend to become horizontally dilated in the northeasterly direction,[1] *i.e., a "delta" pattern will develop within the generally divergent flow on isobaric surfaces.* In these motions inertial effects must be considered, i.e., ageostrophic wind components will develop, and the trajectories of the air particles will cross the contour lines in the direction of lower geopotential heights.

It should be noted that the crest of the waves in the cross-barrier flow will not remain directly over the mountain summit but move some distance downstream, because of inertial overshoot and the generally eastward motion of the planetary waves. Hence, as shown schematically in Fig. 33-b, *the orographically induced diffluent pattern will be advected to the lee side.* On a much smaller scale, such inertial motions and disturbances may become visible in the plume of a banner cloud. If the air mass is fairly stable, various types of gravity waves may also occur in the flow down-stream from the barrier.

When the flow crosses the summit of a mountain, *it will tend to converge in a horizontal plane, while being stretched vertically, because of the increasing depth of the air column.* The flow will attain cyclonic curvature, and *the instantaneous speeds of the particles may be increased due to the horizontal convergence effect.* Therefore, an initially straight and parallel flow crossing a barrier will be deformed not only in the vertical, but also in the horizontal plane as shown in Fig. 33-b.

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[1]However, this idea of the northward displacement of an air flow over the windward side may be compared to the studies done by Hess (1959, see page 252), Kasahara (1966), and others.

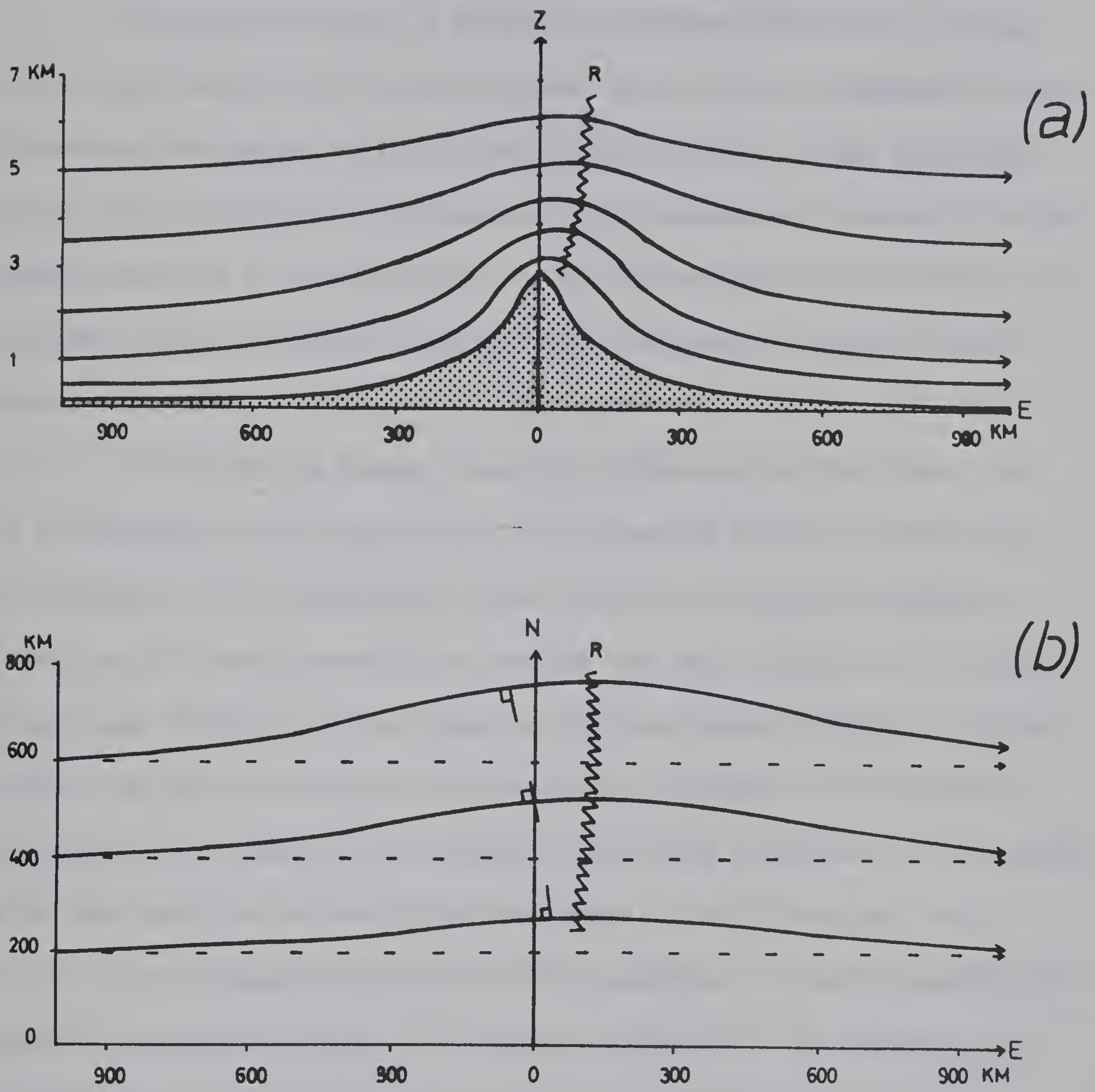

Fig. 33. Showing orographic modification of a uniform airflow over a broad mountain system (north-south extended). (a) Vertical profile with a model mountain of elevation 3 km and half-width 100 km. A ridge R in the airflow appears after the mountain crest. (b) A corresponding horizontal projection of a mid-tropospheric level (say 500 mb). The flow (solid streamlines) has the characteristics of divergence (associated with a diffluent pattern) over the mountain and the relevant ridge of the flow occurs over the lee side. Short dashed lines represent the undisturbed air flow in the absence of a barrier.

5.2 Comments on the Observed Weakening of the 500-mb Flow Over the Canadian Cordillera

As is well known, a layer of incompressible fluid passing over a long barrier will maintain mass continuity by compensating for decreasing thickness with an increase in velocity on the upstream face, and for increasing thickness with a decrease in velocity on the downstream face of the barrier. In a compressible fluid such as the atmosphere the divergence is not zero, in general, and the flow is more complicated.

By way of an example and with reference to Fig. 34-a, let it be assumed that a deep flow of air from the Pacific crosses the Cordillera, with considerable speed and for an extended period of time, say 24 hours, steadily ascending the many ridges of the massif, whose mean slope, from the coast to the Continental Divide, is about 1:200. As the air ascends the Cordillera, large-scale horizontal divergence will ensue as the depth of the flow decreases, in accordance with the equation of Sutcliffe and Godart. Such divergence must occur at all pressure levels where the gradient of the orographically induced vertical velocities ($\partial\omega/\partial p$) is negative. The average orographic vertical velocities are, of course, much smaller than the horizontal wind components, considering the gentle 1:200 mean slope of the Cordillera, but they are still important, and comparable to the vertical motions induced by a warm front of similar slope and aspect. It is also important to note that orographically induced divergence will not be distributed only in the direction of the flow, (i.e., toward the northeast), but also in the transverse direction, parallel to the several ranges, toward the northwest and southeast. *In other*

words, the atmospheric flow is not limited only to a deformation scheme of vertical contraction and dilation or shrinking and stretching.

Frictional effects will reduce the flow speeds over the Cordillera, but it is known that significant reductions in speed do occur only in a relatively shallow boundary layer. Thus, neglecting friction, as the depth of the flow decreases eastward from the Pacific coast, the speed of a *non-divergent flow* at a given level should increase over the Cordillera and reach a maximum over the Divide. However, in the course of this study it was observed that *the horizontal wind does not increase appreciably over the massif, especially not on the 500-mb level; this, it seems, is clear evidence of large-scale horizontal divergence.* More specific evidence was obtained during the examination of the individual upper air patterns associated with the 146 lee cyclones. In almost all cases, the usually southwesterly 500-mb wind field was not noticeably strengthened over the Cordillera. Instead, distinct diffluent patterns were found which tended to fan out appreciably over the Rocky Mountains.

Conversely, jet streams at the 500-mb level usually are found to strengthen slowly beyond the Divide, as the flow converges over the level terrain of the Prairies. Because of the paucity of upper air observations in this mountainous area, there is clearly room for argument, though it is worth noting that *most jet streams attain their maximum intensities somewhere over Eastern North America.* This is clearly shown on the mean maps of jet stream strength and position, published by Namias and Clapp (1949) and also by Petterssen (1956). In addition, there is the supporting evidence that *anticyclones tend to build on the windward side, and that frequently jet streams "split"*

over the Cordillera.[1]

On the basis of the vertical shrinking effect alone, the wind speeds at lower levels near mountain peaks may be expected to increase to some degree. It has been noted that wind speeds increase rather sharply over hills and small mountains, but air particles cross such small obstacles in a relatively *short time*. However, if the mountains are high and extensive, then *the orographic vertical motions over the windward side of the massif produce an overall horizontal divergence which, in time, leads to the establishment of a distinct, diffluent pattern of the kind frequently found over the Canadian Rockies.*

5.3 Observations and Speculative Comments on Orographic Motions and Lee Cyclogenesis

In the following discussion, reference is made to Fig. 34 (b) and (c). As described earlier, an on-shore flow will generally tend to curve to the north, when it begins to ascend the massif. Moreover, the upper flow tends to slow down, as the diffluent pattern intensifies, i.e., fans out north and south, and lengthens west to east. In other words, while a steady-state, diffluent pattern may readily persist over flat terrain, the pattern over the Canadian Cordillera will change with time and grow in all directions. As shown schematically in Fig. 34-b, the result is that the leading edge of the diffluent pattern progresses toward and beyond the Divide; the upstream trough tends to retrograde; and the diffluent pattern as a whole broadens and spreads laterally, to the north and south. *Thus, an important consequence of large-scale divergence appears to be deceleration of the upstream*

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[1]See Reiter (1963), p. 383.

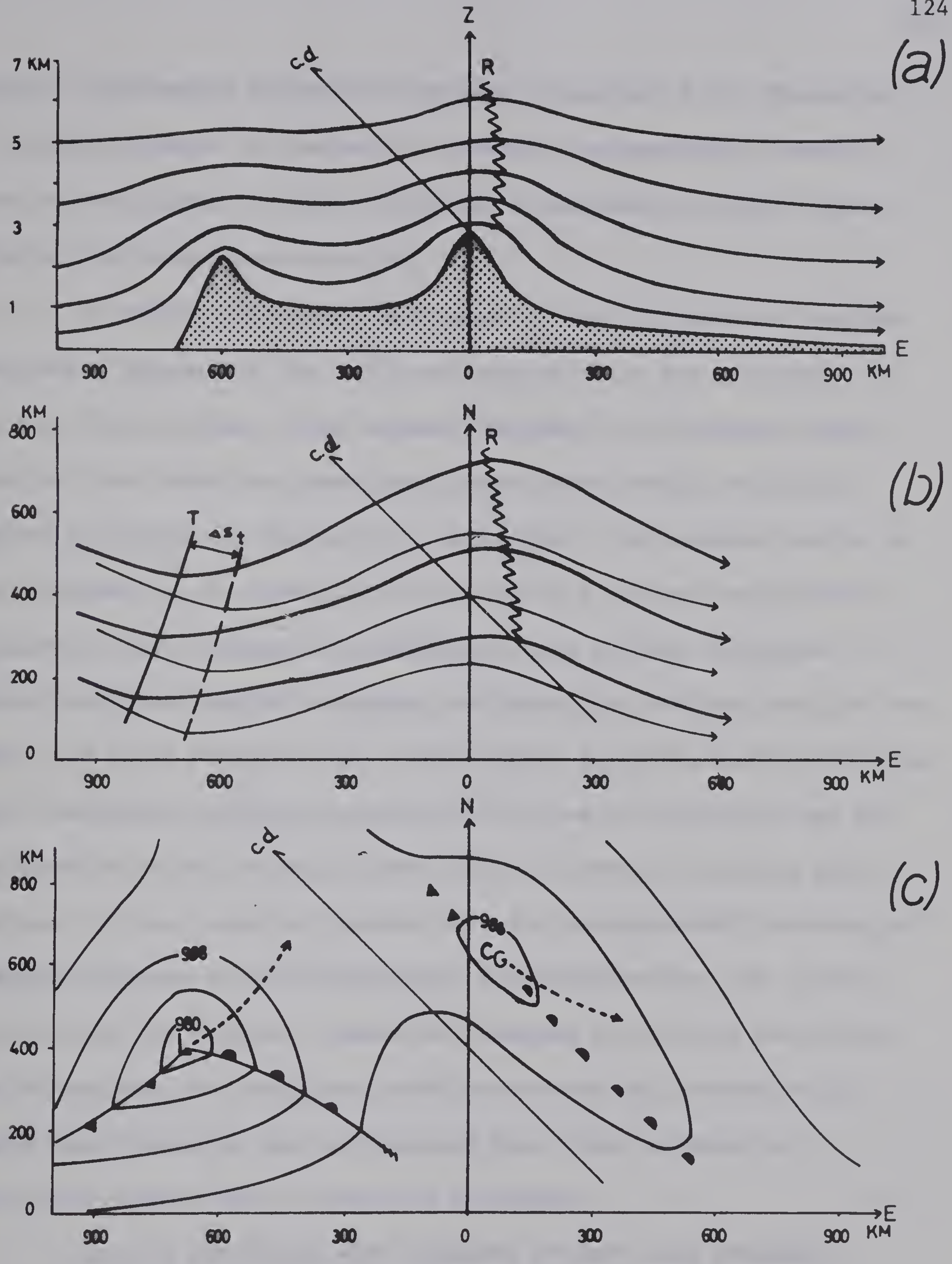

Fig. 34. Illustrating cyclogenesis under an orographically intensified diffluent upper air pattern in the lee of the Canadian Rockies. (a) Vertical profile of the Cordillera. (b) Composite map of two types of upper flow. The flow without mountains is shown by thin streamlines; t is the upper trough, and an upper ridge lies along the N-axis. Flow over the mountains is depicted by thick streamlines which tend to become normal to the Divide, c.d.; T is the upper trough (weakening and decelerating) and R the upper ridge. Δs is the lag between the troughs. (c) The corresponding sea-level chart: Cyclogenesis occurs along the lee slope of the Divide, hundreds of kilometers ahead of the parent surface low, L. Dashed lines indicate trends of system motions.

trough, a phenomenon referred to earlier in section 3.9. The motion of incoming troughs is frequently retarded, and sometimes arrested over the Cordillera, or just off-shore; occasionally, slight retrogression has been observed as well.

A process of crucial importance in lee cyclogenesis involves the leading segment of the diffluent pattern which has progressed to the lee of the Divide. *This leeward "overhang" of divergence will "suck up" and remove air from the columns below, which, being in regions of orographic descent, are subsiding.* The combined action of the processes of air removal from the top of a column, and inherent subsidence (with orographic convergence) will produce alternate regions with substantial ascending and descending motions over the lee slope; and *these motions in turn will induce large vertical stretching.* Such a horizontal diffluent pattern in the lee of the Divide may be considered with the vertical stretching of layers of mountain wave profiles. At any rate, *it appears that the orographically intensified diffluent pattern in mid-tropospheric levels determines the initial formation of lee cyclones.* Moreover, changes in the flow pattern at the 500-mb level are much more readily detected and indicative of significant impending lee cyclogenesis than other evidence of structural changes such as vertical shrinking.

East of the Divide the diffluent pattern will gradually close up and converge again as the depth of the air columns increases. It seems that the vertical stretching effect alone cannot intensify jet streams in the lee. In other words, if there are no horizontal converging motions at upper levels, the jet stream may be weakened in the lee by the stretching of air columns.

This clearly indicates that structural changes in the horizontal plane are as important as changes in the vertical columns (vertical stretching) of air over large mountains. This may lead to the re-intensification of the jet stream in the lee side of the massif and in the eastern sections of the continent, as shown on the mean maps by Namias and Clapp (1949).

When the major part of the diffluent pattern and the upper trough cross the Divide, the behaviour of the lee cyclone on the surface will be governed by, or at least strongly influenced by, the upper system. If the upper system does not cross to the lee side, the local cyclone will remain quasi-stationary and may dissipate. *The horizontal convergence effect will also cause the system to accelerate, as it will be seen from Fig. 35. In turn, the upper wave will begin to accelerate*, as is evident from the movement of troughs at the 500-mb level, discussed in section 3.9. However, Hage (1957) noted that cold lows tended to accelerate when the inflection point of the upper wave passed over British Columbia.

It may be also noticed that *there is a tendency for an upper cold low to become a trough over the windward side, when the diffluent pattern intensifies. For the same reason, a confluent pattern (west of the trough) is weakened by orographic divergence. Hence, the curvature of an upper cold trough decreases and the trough seems to fill (weakening of upper wave) as the contours straighten out.* Consequently, in some of the cases discussed in section 3.9, the upper troughs all but dissipated over the mountains, or at least could not be clearly identified, but they frequently regenerated on the lee side.

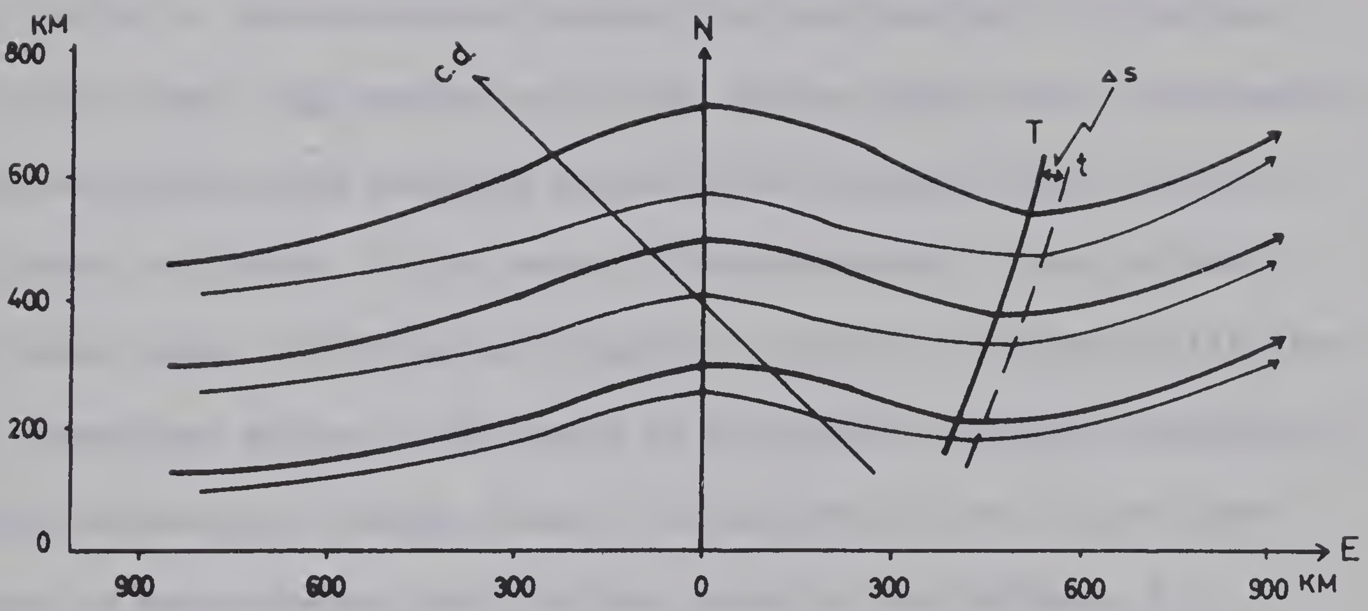

Fig. 35. Plan view illustrating development and acceleration of an upper trough after crossing the Continental Divide. Symbols are as for Fig. 34-b.

Though there must be other effects, such as land and sea interaction, persisting highs on the windward slopes, the speed of the systems, diabatic and adiabatic heating, etc., which influence the motion of pressure systems, most of the cyclones seem to move primarily in accordance with the potential vorticity equation (11) of Rossby. In general, it is found that cyclonic systems, both at the surface and aloft, move abruptly northwards when they approach the Western Cordillera; in the lee, however, they travel along more gently-curved tracks, and in a predominently easterly to southeasterly direction. In the first case, the systems will move in the direction of decreasing relative vorticity with reducing depth of air column; hence, cyclone tracks curve northward to the west of the mountains. Lee cyclones, on the other hand, follow paths of gently cyclonic curvature while they are in down-slope motion. The cause of the sudden northward motions is not fully understood, though clearly the windward slope of the Coast Mountains is much steeper than the lee slope of the Rockies, a dynamically significant difference. In the Gulf of Alaska, land and sea, latitude, and occlusion effects may be also important. It should be noted, however, that lee cyclones may move very *erratically* in the very early stages of their life cycle, before they are, as it were, steered by, and under the control of the upper flow. At any rate, *it was observed that lee cyclone movement in the earlier stages seemed to depend strongly on the motion of the orographic diffluent upper pattern, at least before the main vortices crossed the Continental Divide.*

Furthermore, strong descent and appreciable convergence in the lee of the Rockies will produce a dynamically unstable atmosphere.

The orographically produced convergence cannot reach the main cyclonic system somewhere far upstream, because of the presence of the Cordillera. Under these conditions a separate development system is initiated in due course, and maintained independently under the orographically intensified diffluent pattern with a large separation distance. These features are shown schematically in Fig. 34-c. The various processes tend to enhance each other and produce a "hyper-baroclinic zone" along the steeper part of the lee slope where strong adiabatic motions usually exist as well.[1] As a result of large shear motions and mixing, some of the meteorological elements will not vary much with horizontal distance and finding frontal zones on synoptic maps may be rather difficult in practice. But it was observed that the frontal structure will become better defined and easier to locate as the systems move out of the strong mixing zone. This difficulty of locating the baroclinic zone in the early stages of lee cyclogenesis may account for the observation made in section 3.7 that about half of the present sample cyclones were apparently not initiated on definite fronts. In addition, Petterssen and Smebye (1971) have recently given further evidence on the above, stating that "fronts may or may not be present" with cyclogenesis in North America. However, augmentation of the baroclinicity seems to be necessary for the intensification of cyclones, as no migratory extra-tropical cyclones are observed to attain maximum intensity without the association of a front.

It is of interest to consider a result obtained by Godson (1948) and used by Palmén and Newton (1969) who show that even a modest descending motion of ω_o = 1 μb/sec (≃ -1 cm/sec) would result in a

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[1] This zone is sometimes referred to as a "pseudo-front". See Carlson (1961).

pressure fall of 3.6 mb/hr. Palmén and Newton concluded that horizontal divergence is not necessary for the formation of lee cyclones in such circumstances, although it is an important term in the Margules-Bjerknes equation, and a governing factor in the development mechanism of all "classical" theories discussed earlier. Since forced descent produces a large amount of low-level convergence, they suggested that orographic motions alone may be sufficient to generate a lee cyclone. In the present investigation the calculated strong orographic descending motions were found to be in the range of 3-7 μb/sec near the foothills, as shown in Figs. 27, 28 and 32, but it is possible that such large motions are partly consumed by frictional stresses. For pure ascending motions in a developing cyclone on the polar front, Palmén and Newton (1969)[1] give typical vertical velocities in the middle troposphere of about 6-8 μb/sec (ascent). Thus the numerical values of terrain induced vertical motions are of the same order as the vertical motions in a deepening frontal cyclone. Nevertheless, strong orographic descents alone are probably unable to create a cyclonic lee disturbance in the manner suggested in the previous discussions.

According to McClain (1960) and Krishnamurti (1968), strong convergence occurs usually only at *relatively low levels* of the lee slope. They found that orographic vertical motions are largest at 900-850 mb, but relatively small at the 700-mb level. This is interpreted to mean that all strong downslope motions will result in rapid low level convergence especially near the foothills. In the absence of adequate compensation by horizontal convergence into the

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[1]See page 382.

stretching column of subsiding air, local "anomalous" or "fictitious" pressure falls may occur. A reverse effect, a fictitious pressure rise may appear on the windward side of a barrier, when the air columns undergo differential shrinking without being fully compensated by horizontal divergence. Such localized anomalous changes have been noted frequently in the reported pressures from the Himalayas, in the course of personal experience with synoptic analysis in East Asia.

The rate of intensification of a diffluent upper pattern induced by orography is also important in lee cyclogenesis, but a meaningful study of this effect would require much more frequent and detailed upper air data over the Cordillera and the Eastern Pacific. Any useful computation of the intensification rate would have to consider a number of complex processes such as thermal advection, stability, adiabatic and diabatic effects, the rate of orographic ascent, frictional drag and synoptic-scale mountain waves. Moreover, as is well known, even the divergence in an intense cyclone is numerically small[1], and it is clear, therefore, that orographically-induced upper divergence present during lee cyclogenesis will be even smaller.

The numerical values of orographic ascent on the windward side are probably somewhat larger in magnitude than the computed values of descending motion plotted in Figs. 27, 28, and 32, since the upslope motions on the windward side usually exceed the downslope motions in the lee. Though vertical motions may be also modified by frictional shear, etc., these relatively large ascending motions will intensify a diffluent pattern aloft at roughly the rate of ω_o of those figures.

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[1]See Petterssen (1956) page 293.

A crude estimation indicates that the leeward part of three 500-mb diffluent patterns over the Rockies, which were associated with three cases of lee cyclogenesis, was laterally shrinking at the rate of about 10-15 per cent in 24 hours. It seems reasonable to assume that the rate of intensification of the diffluent pattern by orographic vertical motions on the windward side of the massif is of the same order.

Some authors, in particular McClain (1960) and Schallert (1962), when evaluating mean divergence values over mountain barriers, assumed that the orographically induced motions will vanish at about 3 km above the terrain, since the isothermal pattern did not seem to change noticeably at 500-mb (see also Hess and Wagner, 1948). But judging by the intensification of 500-mb diffluent patterns observed in the present study, it seems the assumption is not adequate for the Canadian Cordillera. No alternative, practical level is proposed at this time, though some mountain-wave profiles suggest also that the level cannot be in the middle troposphere.

An important characteristic of the anticyclonic ridge in mid-tropospheric levels is that it is more stable than other motion systems at comparable levels, but this does not mean that anticyclonic flow is necessarily more stable. Prolonged cross-barrier flow without eddies, though kinematically possible, is dynamically unstable even under anticyclonic flow conditions. That mountains can promote the generation of eddies of all sizes, as well as leeward convergence, was pointed out by Bjerknes as early as 1911, and mentioned before in section 2.4.1. The role of dynamically produced eddies in lee cyclogenesis is imperfectly understood, but it is not inconceivable that they may act as "triggers" which release kinetic energy into pre-existing

baroclinic zones, and thus initiate development of a classical wave cyclone.

In recent work, Petterssen et al. (1962, 1971) distinguished at least two different mechanisms that will "*initiate and support the development*" of extratropical cyclones: Type A[1], the classical frontal wave, thought to be initiated by baroclinic instability; and Type B, "initiated by a finite disturbance in the upper troposphere, rather than by an amplifying perturbation of a frontal zone at low levels". Type B cyclogenesis was first identified in connection with cyclones which developed in the lee of the Rocky Mountains.

In order to avoid any possible misunderstanding in the discussion that follows, it is necessary at this point to quote the most recent formulation of Petterssen's development hypothesis (Petterssen and Smebye, 1971):

> *Cyclone development at low levels commences when and where an area of positive vorticity advection in the upper troposphere (normally on the foreward side of an advancing trough) begins to spread over an area of warm advection (or near-absence of cold advection) in the lower troposphere.*

Petterssen tested this hypothesis on 97 cases of cyclone development and found that only 8 cases "developed without the criterion of the hypothesis being satisfied". Schallert, studying sea-level cyclogenesis, found on the other hand that only 31 cyclones out of 71 cases obeyed the Petterssen hypothesis. The tests were limited to the *prediction of the initial "development"* of cyclones in North America *to the east of the Rocky Mountains.*

Before considering the results of the present investigation in the light of Petterssen's most recent work, it should be noted that

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[1]Types A and B refer here to Petterssen's own classification.

no attempt was made to differentiate between developments of Petterssen's Type A and B. Moreover, the highest level examined was the 500-mb level, i.e., the mid-troposphere.

It is not clear whether, and to what extent, the present results are at variance with Petterssen's, for much seems to depend on the meaning ascribed to the term "*development*". When Petterssen speaks of "initial development", and "the initiation and support of development", he seems to mean cyclogenesis, in the first case, and cyclogenesis followed by intensification and growth, in the second. And while cyclone initiation and growth certainly do occur as specified by Petterssen's hypothesis, it is equally certain that lee cyclones are frequently initiated beneath the "canopy" of the diffluent pattern, without the presence of a finite upper disturbance or the benefit of positive vorticity advection at the 500-mb level.

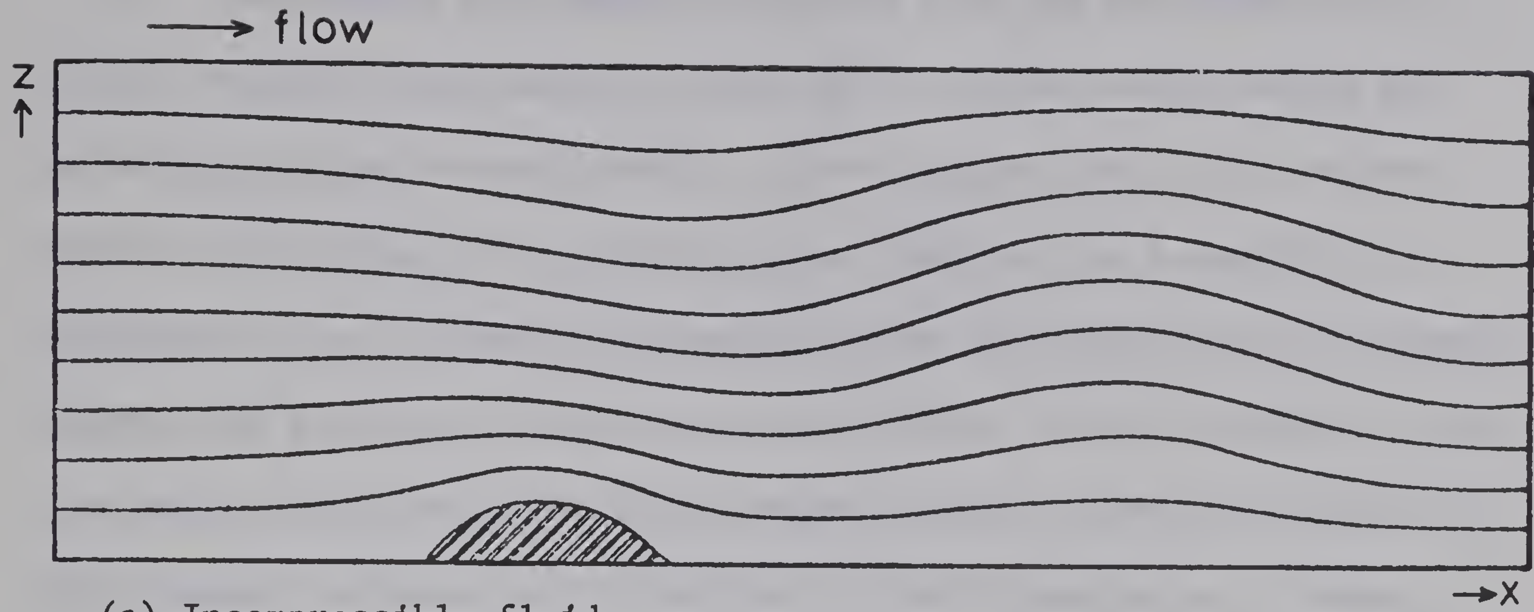
However, if "development" means simply growth and intensification of a small, pre-existing lee cyclone, in accordance with the established criteria of Petterssen's hypothesis, then any differences, real or apparent, are resolved, for many of the cyclones in the present sample intensified in response to pronounced vorticity advection aloft. Moreover, it is often not possible to identify two clear-cut phases of cyclogenesis and intensification, for the evolution of the cyclone may proceed smoothly from birth to rapid growth without a distinct break or pause.

Due to the complexities of three-dimensional modelling and computation, most published lee wave profiles show only vertical sections with all the limitations inherent in two-dimensional models. Frequently, the amplitude of waves is simply proportional to the

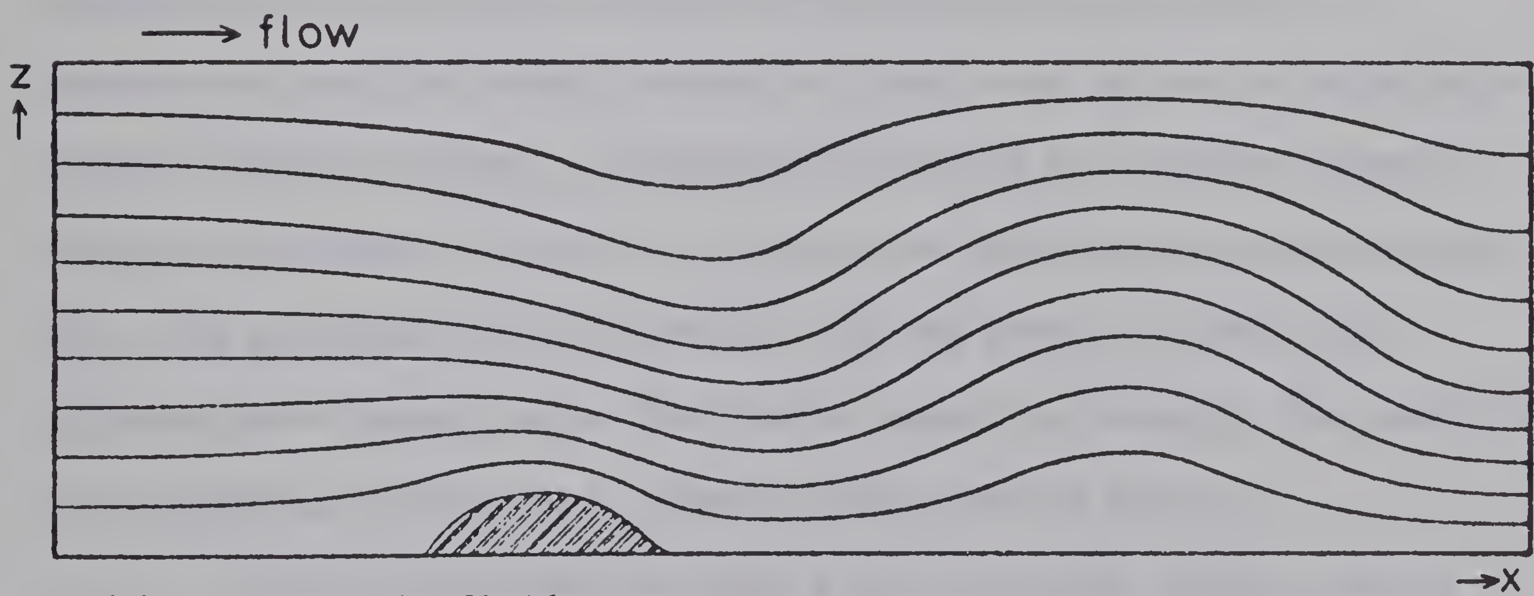
mountain height, and important features of the atmospheric flow over the "real" mountains are modelled inadequately. Thus if mountain passes and finite blocks of north-south extended mountains are considered together with the added complication of non-uniform winds in horizontal planes, the profiles would be shaped very differently.

An interesting, yet still overly simple example of barrier flow, found in a theoretical study by Claus (1964), shows striking differences in the wave characteristics developed in an incompressible and a compressible fluid. As shown in Fig. 36, downstream from the obstacle the amplitude of the wave in the lower region of the compressible flow ("lower atmosphere") is about three times as large as the corresponding vertical oscillation in the incompressible flow, under similar initial and boundary conditions of fluid motion. Moreover, the characteristics of atmospheric divergence over mountains cannot be satisfactorily modelled in liquids, and the shear motions in an incompressible fluid are usually much smaller. It seems reasonable, therefore, to conclude that wave profiles, derived on the assumption of incompressible flow, do not adequately represent the complicated phenomenon of atmospheric barrier flow.

Most mountain wave profiles show vertical velocity distributions which do suggest ways whereby a cyclone may be generated under the vertically stretching layer. If a cyclone travels along a system of waves, such as of a Queney profile, one may expect it to dissipate under vertically shrinking layers which are usually present somewhere east of the Divide. In the present study, it is not clear whether many cases of cyclolysis were caused by this effect, but it was observed that many lee cyclones weakened considerably during at least part of the time while travelling eastward.



(a) Incompressible fluid.



(b) Compressible fluid.

Fig. 36. Comparison of lee waves produced under conditions of incompressible and compressible fluid flow. After Claus (1964).

5.4 Orographic Energy Conversion

Petterssen and Smebye have shown that the net generation of kinetic energy occurs mainly at the 500 to 100-mb levels during the period of cyclone intensification. They suggest that a jet stream exports this energy to the layer below. However, in a general atmospheric flow, there will always be some interconversion of internal, kinetic and potential energy, especially along frontal surfaces. Since a mountain slope has some of the properties of a frontal surface, the total energy exchange may be similar to that occurring at a front. Kinetic energy transformed orographically into potential energy will add to the total potential and internal energy of the system; there will be also some reduction in the kinetic energy by frictional drag, precipitation, etc.

If the system arrives over the lee slope, a large part of the orographically increased potential and internal energy will be reconverted into the kinetic energy of down slope motion and subsidence of the diffluent system. A substantial part of the released kinetic energy will become available to initiate and maintain lee cyclogenesis. Thus, the principal source of energy for the growth of *local* lee cyclones would appear to be the kinetic energy released by the conversion of orographically increased potential and internal energy.

It is often found that when a cold air mass invades the lee of the Rockies from the windward side, a cyclogenetic system deepens much more rapidly. In developments of this nature, the activity of the cold front is apparently increased as a result of the rapid conversion of potential to kinetic energy, as the cold air mass moves down the lee slope of the Rocky Mountains. But when warm air is over-

running cold air on the lee slope, the rate of development is somewhat slower.

5.5 Thermal Lows

On hot days, surface cyclones appeared on a few occasions (e.g., cases 52 and 62) in the Washington-Oregon basin. These so-called "heat lows" are of thermal origin and initially frontless, but they may later develop characteristics associated with frontal waves. In the early stages of their development, these cyclones migrate from their source region essentially in the direction of the diffluent upper pattern, through the interior of B.C. and sometimes as far as the Southwestern Alberta Range. The development of heat lows seems to proceed as follows: at first, a severe local heating effect produces a large amount of low level convergence and induced vertical motions; secondly, an orographically intensified diffluent pattern over the Cascade Range and the Coast Mountains further increases the ascending motion within the disturbance due to the "sucking effect" of mid-level divergence. This mechanism satisfies the basic requirements of the compensation theorem of vergence for cyclone development in the troposphere. Weak cyclones form often also in the interior valleys of B.C., but they soon dissipate when they meet the main range of the Rockies.

Convective clouds are often reported over the lee slopes of the Canadian Rockies when an orographically induced diffluent flow pattern is present aloft. It seems that such meso-scale and lesser synoptic-scale disturbances develop more readily when the net divergence in the middle troposphere exceeds the low level convergence caused by the terrain.

5.6 Summary and Conclusions

According to Petterssen there appear to be at least two main types of extratropical cyclone development : First, Type A, the frontal wave cyclone, initiated by low-level baroclinic instability, and intensified largely by *thermal* advection. This is, of course, the classical amplifying wave cyclone of the Bergen school of meteorologists led by Bjerknes, Bergeron and Solberg. Second, the Type B cyclone, induced by a finite disturbance in the upper troposphere, initiated and sustained primarily by strong *vorticity* advection aloft. This is the vorticity-induced cyclone of the Chicago school, first identified by Petterssen, and studied by many authors. Many cyclones are neither of pure Type A nor of Type B, but a combination of the two. This is particularly true of the intense lee developments described and classified in this study as Type A.

In addition to the major developing cyclones, there are various types of small local cyclones, usually weak and non-developing systems (Types B, C and D) which occur in the lee of large mountain barriers. These are generated by purely orographic mechanisms, or by other means, e.g., heat lows, influenced to some extent by orography, but largely products of intense solar heating. The cyclone classification used here is based on Schallert's 1962 nomenclature, and not fully compatible with Petterssen's. However, the results of this investigation agree in the main with the findings of Petterssen, Hage, and Schallert, insofar as they apply to cyclone development in the lee of the Canadian Rockies.

With the upper airflow at a large angle to the principal ranges a zone of strong baroclinicity usually appears in the lee of the

Canadian Rocky Mountains. The cross-barrier flow frequently leads also to the development of a surface lee trough, a geographically bound source region of cyclogenesis. The most striking feature is that cyclogenesis is commonly initiated in the lee of the Rockies under conditions of marked divergence aloft, which is characterized by a diffluent, cross-barrier flow pattern at the 500-mb level. Bjerknes showed that surface cyclones develop approximately below the inflection point, i.e., about half way between the trough and the next downstream ridge of the upper wave, but it appears that, in the case of lee cyclogenesis, the site of initiation is shifted well toward the diffluent flow of the downstream ridge. Cyclogenesis may also result from vorticity advection in the middle and upper troposphere, but the basic mechanism responsible for the initiation of most lee cyclones, particularly of feeble and non-developing systems, seems to be mid-level divergence. However, sudden intensification of *mobile cyclones* usually commences with the onset of positive vorticity advection, in accordance with Petterssen's hypothesis.

Detailed analysis indicates that the frequency maximum of lee cyclogenesis in the Canadian Rockies is not single, but composed of three distinct centres. The main centre of activity is located in the lee of the Southwestern Alberta Range, while two secondary maxima are found in the lee of the Northern B.C. Range and the Mackenzie Mountains. The frequency maxima appear about 200-250 km from the Continental Divide. This distance could be identified with a region of vertical stretching and, perhaps, be a parameter important in synoptic-scale mountain wave theories, but the evidence so far is inconclusive; no reference to lee wave studies directly concerned with

lee cyclogenesis has been found in the literature.

It has been noted that many low pressure systems tend to move northward on the windward slope, and generally southeastward in the lee of the Cordillera, though many incipient lee cyclones do remain quasi-stationary when the motion of the (orographically intensified) upper diffluent pattern is small or quasi-stationary. When surface lows decay on the windward side, the associated upper wave systems are usually found to have been weakened over the Cordillera. On the other hand, once past the Divide, the upper winds, and especially jet streams, tend to strengthen over the lee slope, and usually continue to intensify over Eastern Canada and the Northeastern U.S.

Many numerical prediction models incorporate what appear to be still overly smoothed terrain profiles, especially near the Continental Divide; it seems that more realistic profiles of the Canadian Rockies, of proper 2-3 km height along the Divide, would yield better results. Orographic vertical motions were found to be quite large on the lee slope and very important in the generation of lee disturbances. Accurate computation of the orographic vertical velocity field would seem impossible in view of the great complexity of the actual terrain, but good estimates obtained from smoothed topographic profiles are probably adequate for most synoptic purposes. However, depending on the size and slope of the orographic massif, a diffluent pattern could be modified considerably by vertical motions, as may be readily deduced from the Sutcliffe-Godart continuity equation. It is highly plausible that orographically induced divergent motion over the windward slope is compensated by vertical stretching and horizontal shrinking over the lee slope. Incomplete or lagging compensation ("imbalance compensation")

may well be the principal mechanism which initiates not only the weak, local cyclones but also the nascent cells of low pressure that develop into intense, migrating cyclones. Further research on lee cyclogenesis should concentrate on the question of orographic intensification and weakening of diffluent upper-air wind fields over the Canadian Cordillera, but realistic computations of vertical velocities and upper divergence may have to await the establishment of a denser network of RAWIN stations, or at least the soon to be hoped-for temperature and moisture profiles deduced from satellite radiation data.

The results of this investigation may be somewhat at variance with a number of earlier studies which *appear* to conclude that positive vorticity advection is necessary for the *initiation* of lee cyclones. According to Hage (1957) and Petterssen and Smebye (1971), lee cyclones *develop* when upper vorticity advection "*begins to spread*" over a region of low level fronts or of warm advection. This would seem to be an overly restrictive requirement, meant to apply apparently only to moderate and intense, developing cyclones.

In the course of the present analysis it was observed that *most lee cyclones formed initially under the eastern margin of a 500-mb, orographically intensified, diffluent cross-barrier flow, superimposed on a zone of low level convergence and orographic descent.* But *sudden deepening* of lee cyclones was found to be associated with upper cold troughs and positive vorticity advection, as required by Petterssen's hypothesis. In the development of pressure systems over mountains, a divergence field is predominant over the windward side which, accompanied by orographic ascent, produces a cell of divergence in the

lower and mid-troposphere, and a compensating cell of convergence in the stratosphere and upper troposphere. In the lee of the range, the cyclogenetic fields are generated and compensated approximately in the converse manner necessary for the development of extratropical cyclones, as postulated by Dines, Scherhag, J. Bjerknes and Sutcliffe.

LIST OF SYMBOLS

A_Q	$= -\vec{V}\cdot\nabla Q$, vorticity advection at the level of non-divergence
A_T	$= -(u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y})$, temperature or thickness advection
c_p	specific heat of dry air at constant pressure
D_p, D	isobaric divergence and horizontal divergence, respectively
$\frac{dp}{dt}$	$= \omega$, vertical velocity expressed in terms of pressure
$(\frac{\partial p}{\partial t})_z$	local change of pressure with time at the level z
$\frac{dW}{dt}$	heat energy given to a unit mass per unit time
F	vorticity of frictional force
f	$= 2\Omega\sin\phi$, the Coriolis parameter
g	acceleration of gravity
H	$= \log(\frac{p_o}{p})\frac{1}{c_p}\frac{dW}{dt}$, thickness change due to diabatic heating
n	a measure of length normal to isentropic surfaces
p_o, p	pressure at the lower and upper level, respectively
Q	absolute vorticity
Q_o	absolute vorticity at sea-level
q	relative vorticity
R	gas constant for dry air
γ	dry adiabatic lapse rate of air temperature $-\frac{\partial T}{\partial z}$
ρ	density
S	$= \log(\frac{p_o}{p})\frac{dp}{dt}(\frac{r}{g\rho} - \frac{\partial T}{\partial p}) = \log(\frac{p_o}{p})\overline{\omega(\Gamma_a - \Gamma)}$, the buoyancy term
T	air temperature
t	time
θ	potential temperature

$\bar{h}$	thickness between two isobaric surfaces
$\vec{V}$	wind vector
w	vertical component of wind
w_n	component of velocity normal to isentropic surface
z	vertical space coordinate
∇	two-dimensional differential operator
∇^2	two-dimensional Laplacian operator
∇_3	three-dimensional differential operator

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APPENDIX

Table 2. List of cases of lee cyclones in monthly order (1958)

Case number	Date	Source region	Development type
1	00Z/01/JAN - 12Z/03/JAN	Alberta	B
2	00Z/04/ " - 00Z/06/ "	Mackenzie	A_1
3	00Z/08/ " - 12Z/09 "	Mackenzie	A_1
4	00Z/09/ " - 00Z/11/ "	Alberta	A_1
5	00Z/11/ " - 00Z/13/ "	"	A_2
6	18Z/12/ " - 00Z/14/ "	"	A_1
7	00Z/15/ " - 18Z/15/ "	"	D
8	18Z/15/ " - 18Z/16/ "	"	D
9	06Z/17/ " - 06Z/18/ "	"	D
10	18Z/17/ " - 12Z/18/ "	Mackenzie	D
11	18Z/19/ " - 12Z/21/ "	"	D
12	00Z/21/ " - 12Z/22/ "	Alberta	C
13	18Z/23/ " - 12Z/26/ "	"	C
14	12Z/24/ " - 18Z/27/ "	Northeastern B.C.	D
15	18Z/28/ " - 06Z/30/ "	Alberta	A_1
16	12Z/30/ " - 00Z/01/FEB	"	D
17	18Z/01/FEB - 00Z/05/ "	Northeastern B.C.	A_2
18	18Z/17/ " - 12Z/19/ "	Mackenzie	C
19	12Z/19/ " - 00Z/22/ "	Alberta	A_1
20	12Z/22/ " - 18Z/23/ "	Northern Alberta[1]	C
21	12Z/23/ " - 18Z/25/ "	Alberta	A_2
22	00Z/25/ " - 00Z/27/ "	"	C
23	12Z/01/Mar - 18Z/03/MAR	Northern Alberta	A_1
24	12Z/03/ " - 06Z/06/ "	"	C
25	18Z/17/ " - 06Z/19/ "	"	D
26	06Z/21/ " - 18Z/22/ "	"	A_2
27	06Z/21/ " - 06Z/22/ "	Alberta	D
28	06Z/25/ " - 18Z/26/ "	"	A_1
29	18Z/29/ " - 18Z/31/ "	Northern Alberta	A_1
30	18Z/30/ " - 12Z/02/APR	Montana	A_2
31	06Z/31/ " - 00Z/03/ "	Alberta	A_2
32	00Z/04/APR - 18Z/04/ "	Mackenzie	C
33	12Z/08/ " - 06Z/10/ "	"	A_1
34	00Z/09/ " - 12Z/11/ "	Alberta	A_2
35	12Z/12/ " - 06Z/14/ "	Mackenzie	C
36	18Z/12/ " - 06Z/15/ "	Northeastern B.C.	A_1
37	00Z/14/ " - 12Z/16/ "	Alberta	A_1
38	12Z/15/ " - 12Z/16/ "	Northeastern B.C.	A_1

[1] Cases which occurred near the northern border of Alberta.

Table 2. Continued:

Case number	Date	Source Region	Development type
39	12Z/16/APR - 00Z/19/APR	Alberta	A_2
40	18Z/16/ " - 12Z/17/ "	Montana	A_2
41	18Z/17/ " - 18Z/18/ "	"	D
42	06Z/18/ " - 00Z/22/ "	Alberta	A_1
43	06Z/18/ " - 18Z/19/ "	Northeastern B.C.	D
44	00Z/20/ " - 00Z/25/ "	Alberta	A_1
45	06Z/21/ " - 00Z/22/ "	"	D
46	00Z/04/MAY - 12Z/05/MAY	"	A_1
47	00Z/09/ " - 12Z/13/ "	Mackenzie	A_1
48	00Z/09/ " - 12Z/11/ "	Saskatchewan	C
49	18Z/11/ " - 18Z/16/ "	Montana	A_2
50	18Z/14/ " - 06Z/22/ "	Mackenzie	A_1
51	06Z/20/ " - 00Z/24/ "	Northeastern B.C.	C
52	12Z/23/ " - 06Z/26/ "	Montana	C
53	12Z/24/ " - 12Z/28/ "	Mackenzie	A_2
54	00Z/27/ " - 06Z/29/ "	"	C
55	00Z/29/ " - 00Z/31/ "	Alberta	D
56	06Z/02/JUN - 12Z/04/JUN	"	C
57	18Z/04/ " - 12Z/08/ "	Mackenzie	A_1
58	06Z/09/ " - 00Z/14/ "	Alberta	A_1
59	18Z/13/ " - 12Z/20/ "	Mackenzie	A_1
60	18Z/14/ " - 18Z/18/ "	Alberta	C
61	00Z/22/ " - 00Z/28/ "	Mackenzie	A_1
62	12Z/25/ " - 00Z/27/ "	Northeastern B.C.	B
63	00Z/29/ " - 06Z/02/JUL	Mackenzie	D
64	18Z/11/JUL - 12Z/15/ "	"	A_2
65	12Z/15/ " - 12Z/19/ "	"	A_1
66	18Z/18/ " - 00Z/23/ "	"	C
67	00Z/21/ " - 18Z/23/ "	"	D
68	12Z/22/ " - 12Z/26/ "	Alberta	A_1
69	06Z/25/ " - 00Z/27/ "	Mackenzie	D
70	12Z/26/ " - 12Z/28/ "	Montana	C
71	06Z/29/ " - 18Z/01/AUG	Mackenzie	B
72	12Z/01/AUG - 18Z/02/ "	Alberta	D
73	12Z/03/ " - 18Z/07/ "	"	A_1
74	18Z/06/ " - 00Z/10/ "	Mackenzie	A_2
75	18Z/07/ " - 00Z/11/ "	Alberta	C
76	06Z/10/ " - 12Z/13/ "	"	A_2
77	18Z/12/ " - 06Z/14/ "	"	D
78	00Z/14/ " - 12Z/19/ "	Mackenzie	A_1
79	06Z/15/ " - 12Z/19/ "	Alberta	D
80	18Z/21/ " - 12Z/25/ "	"	A_1
81	18Z/22/ " - 06Z/26/ "	Northeastern B.C.	D
82	18Z/24/ " - 12Z/28/ "	Mackenzie	A_1
83	00Z/26/ " - 18Z/28/ "	Alberta	A_2
84	18Z/27/ " - 18Z/30/ "	Montana	A_2
85	12Z/28/ " - 00Z/02/SEP	Northern Alberta	A_1
86	06Z/01/SEP - 12Z/04/ "	Alberta	A_2
87	12Z/04/ " - 12Z/07/ "	Mackenzie	A_1
88	06Z/07/ " - 18Z/10/ "	Northeastern B.C.	A_2

Table 2. Continued:

Case number	Date	Source Region	Development type
89	12Z/09/SEP - 06Z/12/SEP	Mackenzie	A_1
90	12Z/09/ " - 00Z/11/ "	Montana	D
91	12Z/12/ " - 18Z/15/ "	Alberta	A_1
92	18Z/16/ " - 00Z/20/ "	Northeastern B.C.	A_1
93	06Z/19/ " - 00Z/23/ "	Alberta	A_2
94	12Z/21/ " - 12Z/22/ "	Northeastern B.C.	B
95	00Z/22/ " - 06Z/26/ "	Montana	A_2
96	00Z/25/ " - 00Z/26/ "	Mackenzie	D
97	00Z/28/ " - 06Z/01/OCT	Alberta	A_2
98	06Z/28/ " - 06Z/01/ "	Mackenzie	C
99	06Z/02/OCT - 18Z/05/ "	"	A_2
100	00Z/03/ " - 18Z/05/ "	"	A_2
101	06Z/03/ " - 12Z/05/ "	"	C
102	12Z/04/ " - 18Z/08/ "	"	C
103	06Z/05/ " - 00Z/07/ "	Alberta	D
104	06Z/07/ " - 12Z/09/ "	Northeastern B.C.	A_2
105	12Z/08/ " - 06Z/11/ "	Saskatchewan	A_1
106	12Z/12/ " - 06Z/15/ "	Alberta	D
107	18Z/12/ " - 00Z/15/ "	Northeastern B.C.	C
108	12Z/14/ " - 00Z/17/ "	Alberta	A_1
109	00Z/19/ " - 12Z/22/ "	Northeastern B.C.	A_1
110	00Z/20/ " - 00Z/25/ "	Montana	A_1
111	06Z/21/ " - 06Z/22/ "	Mackenzie	C
112	00Z/22/ " - 00Z/23/ "	"	C
113	06Z/23/ " - 18Z/25/ "	Northern Alberta	A_1
114	06Z/26/ " - 00Z/28/ "	Mackenzie	C
115	06Z/28/ " - 12Z/29/ "	"	C
116	18Z/29/ " - 16Z/01/NOV	"	A_1
117	00Z/01/NOV - 06Z/03/ "	"	A_1
118	18Z/02/ " - 06Z/06/ "	"	C
119	12Z/04/ " - 06Z/07/ "	Montana	A_1
120	00Z/07/ " - 12Z/10/ "	Alberta	A_1
121	06Z/07/ " - 00Z/11/ "	Mackenzie	C
122	18Z/09/ " - 12Z/10/ "	Northeastern B.C.	D
123	00Z/10/ " - 06Z/12/ "	Montana	A_1
124	18Z/11/ " - 18Z/12/ "	Northern Alberta	D
125	12Z/12/ " - 06Z/15/ "	Alberta	A_1
126	12Z/18/ " - 12Z/19/ "	Mackenzie	D
127	18Z/20/ " - 06Z/22/ "	Alberta	C
128	18Z/21/ " - 00Z/24/ "	"	C
129	00Z/23/ " - 12Z/24/ "	Montana	A_2
130	18Z/24/ " - 12Z/27/ "	"	A_2
131	18Z/27/ " - 06Z/29/ "	Northeastern B.C.	B
132	06Z/30/ " - 18Z/01/DEC	Alberta	C
133	18Z/01/DEC - 12Z/03/ "	"	A_1
134	18Z/02/ " - 00Z/06/ "	"	A_2
135	00Z/12/ " - 00Z/13/ "	"	D
136	18Z/15/ " - 18Z/17/ "	"	A_1
137	00Z/17/ " - 06Z/19/ "	Northern Alberta	A_1

Table 2. Continued:

Case number	Date	Source Region	Development type
138	00Z/18/DEC - 18Z/19/DEC	Alberta	A_2
139	12Z/20/ " - 12Z/21/ "	"	B
140	12Z/21/ " - 06Z/23/ "	Northern Alberta	C
141	18Z/23/ " - 12Z/24/ "	Saskatchewan	D
142	06Z/24/ " - 18Z/28/ "	Northeastern B.C.	C
143	12Z/25/ " - 06Z/26/ "	Saskatchewan	D
144	00Z/26/ " - 00Z/28/ "	Alberta	C
145	18Z/26/ " - 12Z/29/ "	"	A_1
146	12Z/29/ " - 00Z/31/ "	Mackenzie	D